MEDDELANDEN FRÅN UPPSALA UNIVERSITETS GEOGRAFISKA INSTITUTION
Ser. A *Nr. 194*

LAKE VÄTTERN

INVESTIGATIONS ON SHORE AND BOTTOM MORPHOLOGY

BY

JOHN O. NORRMAN

SÄRTRYCK UR
GEOGRAFISKA ANNALER, HÄFTE 1-2, 1964

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INVESTIGATIONS ON SHORE AND BOTTOM MORPHOLOGY

Inaugural Dissertation

BY

JOHN O. NORRMAN

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By due permission of the Faculty of Science of the University of Uppsala, to be publicly discussed in the lecture hall of the Institute of Geography, May 19, 1964, at 10 a.m. for the degree of Doctor of Philosophy.

Esselte Aktiebolag

STOCKHOLM 1964

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BIBLIOTHEEK
LEIDEN

Tryckt med bidrag från Längmanska kulturfonden. Samtliga flygbilder godkända för spridning av Försvarsstaben. Kartskisserna godkända för publicering i Rikets Allmänna Kartverk den 13 april 1964.

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INTRODUCTION

Since the second world war intensive research has been pursued on shore processes and shore morphology. This research has mainly been carried out in the fields of hydraulics, hydraulic engineering, oceanography, submarine geology and physical geography. The urgent need of knowledge about wave forecasting and wave conditions in nearshore waters for shipping and landing operations during the war was undoubtedly of great importance for the initiation of this research. To-day the applied research is concentrated on the problems of beach protection.

During the 1930s an important advance was made in our knowledge of the general principles of erosion, transportation and sedimentation and of sediment analyses, but the morphological studies lagged behind. As regards coast and shore morphology, there was certainly relative stagnation during the 1920s and 30s, which is reflected by the long interval between D. W. Johnson's classical handbook "Shore processes and shoreline development" (1919) and the modern handbooks by Valentin (1952), Guilcher (1954, English ed. 1958) and King (1959). de Martonne pointed out that the geographers owed the engineers much (1926, p. 972), but it would be a long time before there was a really fruitful collaboration between the different branches of research.

Most of the post-war advance as regards shore processes can be traced back to the results of research in the fields of hydraulics, applied hydraulics and sedimentology. Important papers have been produced at several research centres, both in Europe and the United States, but it seems to me that none have been of such general importance as the Memoranda and Reports edited by the Beach Erosion Board, U.S. Army Corps of Engineers. Their importance is well illustrated by King's reference lists. Many of these papers deal with hydraulics, applied hydraulics, and the technical problems of beach protection, but the fundamental problems are the same as are met with in studies of shore processes and their morphological effects.

In Sweden there has not been much research on shore processes and shore morphology, which may be explained by the fact that there has been little abrasion on our coasts in modern times. A rather short but progressive paper on the morphological development of lake shores was published by Ahlmann in 1914. The flat coasts of the southern Baltic were studied by Richter (1936), and the flat coast of Scania has recently been investigated by Davidsson (1963). Raised shorelines have been much used in the studies of the uplift since the deglaciation, but there are only a few studies devoted to the morphology and the development of these shores (cf. Hessland 1943, Järnefors 1949, Gillberg 1952).

The Swedish lakes can be said to be the "empty spaces" in the geological mapping. There have been systematic investigations on the organic sediments (cf. G. Lundqvist 1925, 1942 *b*), but none on the inorganic sediments. Except for a brief description of the shores of Lake Torneträsk (Sjögren 1909), no investigations on the shore and bottom morphology of any of the large Swedish lakes have been published.

My interest in shore processes and shore morphology was roused in 1954 during a field study of shore retreat in a dammed part of the river Indalsälven. This study was initiated by Dr. C. C:zon Caldenius, the late senior geologist of the Swedish Geological Survey, and he also suggested that I should continue my studies in Lake Vättern, where he had made some preliminary observations on the present abrasion (1951).

The aim of the investigations on the shore and bottom morphology of Lake Vättern discussed in this paper was to describe the characteristic shore and bottom features and to analyse these features with regard to the conditions which have determined the present morphology. The first part of the paper deals with the general conditions of the Vättern basin, its geology and geomorphology, climate and hydrology. The second part is a synopsis of the principles of shore processes and shore morphology, and in the third and fourth parts the shore and bottom morphology of the lake are discussed.

Lake Vättern is convenient in several respects for fundamental studies of shore and bottom morphology. With a length of 130 km, a maximum width of 31 km, and an average width of 15 km, it is large enough for the generation of waves, which have an appreciable morphological effect. The development since the deglaciation is problematic, because of the great changes in the water level. It can be estimated that during the early stage of the deglaciation the water level at the southern end of the lake was dammed by the ice to at least 140 m above the present water level, and during the following sea stage of the basin the water level in this area dropped to about 35-40 m below the present level. These changes involve problems of the separation of sediments formed by shore processes during the present and previous conditions, and the possibility of glacifluvial sedimentation must also be taken into account.

For my approach to the problems of the previous conditions the investigations by Granlund (1934 *a*) on the age of submerged lagoon deposits of the period from 2500 B.C. to the present time, and by E. Nilsson (1939, 1953, 1958) on the raised shorelines formed in the period from the deglaciation to about 5700 B.C. have been most valuable.

The Vättern investigations started in the winter of 1955/56, and the first studies were mainly devoted to the shores of Visingsö island (Norrman 1959). During the years 1958 and 1959 most of my time in the field was spent on an investigation on the shore erosion of the river Göta Älv (Sundborg and Norrman 1963). Intensified studies in Lake Vättern were again taken up in 1960, and the main field work ended in 1962. Some additional observations were made as late as the autumn of 1963.

In order to make the course of the investigations comprehensible, the equipment should be mentioned. When the investigations started, there was no other equipment than level, compass and lead line. During the summer of 1956 it was possible to rent a fishing boat and a recording echo-sounder for three weeks. The following year an underwater camera was bought, diving equipment was borrowed from the Royal Swedish Navy, a rowing boat from the Visingsö airfield, and an anemometer from the Meteorological Institute in Uppsala. In 1960 a 34 ft motor boat was bought with private loans, and the Institute of Geography furnished a recording echo-sounder, a dredge and diving equipment. In the following year a davit, a piston core sampler, a propeller log and a tacheometer theodolite were obtained.

A considerable portion of the field studies had to be devoted to preliminary surveys because of the poor information available about the general shore and bottom morphology. A brief manuscript on the morphology of the western mainland shore, written by a student, without any sketch maps or profiles, was found in the library at the Institute of Geography (S. Pehrson 1942), and at my request Professor Hjulström sent a student to make a preliminary survey on the eastern mainland shore (L.-O. Johansson 1955). Under my instructions special investigations were carried out on the south-western shore of Visingsö island (Ella Nilsson 1957) and on the Rosenlund cliffs at the southern end of the lake (P. A. Björklid 1962). In addition, I examined all these areas myself.

The bottom morphology was only known through the chart published in 1879, and through the notes of superficial samples taken by Professor Sven Ekman in 1911 and 1912 for an inventory of the bottom fauna. The occurrence of terminal moraines, glacifluvial deltas and traces of ice oscillations on the surrounding land and the contradictory opinions as regards the ice conditions in the basin during the deglaciation necessitated extensive echo-soundings across and along the lake. Thus the investigations have in part had the character of explorations.

Acknowledgements

This work was made possible by support and assistance from many private persons and institutions. First of all, I want to mention Professor Filip Hjulström, head of the Institute of Geography at the University of Uppsala, who taught me geomorphology in the spirit of his "Uppsala school", and whose general support and help with equipment for the investigations were most valuable. He has also read the proofs. With Dr. Åke Sundborg, head of the Geomorphological Laboratory, I have over the years discussed many problems concerning shore processes and sedimentology, and I have learnt much from him, especially during our investigation on the river Göta Älv. He has read the parts of the manuscript dealing with the climate and the hydrology, the principles of shore processes and the shore investigations. The application of diving techniques to geomorphological studies was discussed with Dr. Lennart Arnborg. Dr. Anders Rapp, editor in charge of the *Geografiska Annaler* during the printing, has read the chapters on the Quaternary geology and the bottom investigations. Dr. Bengt Collini, the Associate Professor at the Institute of Geology, University of Uppsala, has given me valuable information about the bedrock geology of the Vättern basin, partly by field demonstrations, and he has also read the part of the manuscript dealing with this geology. Notes on samples taken in Lake Vättern in 1911 and 1912 were kindly lent to me by the late Professor Sven Ekman. Samples of organic deposits were discussed with Mr. L.-K. Königsson, F.K., and samples of wood were examined by Mr. T. Ingmar, F.M. In discussions many colleagues in the Departments of Physical Geography and Quaternary Geology have made constructive criticisms and valuable suggestions too numerous to specify in detail. All my colleagues cannot be mentioned by name, but I want to make special acknowledgements to Mr. V. Axelsson, F.L., Mr. Å. Hörnsten, F.L., and Mr. T. Stenborg, F.L.

I wish specially to thank the friends who assisted me during the field work: Mr. Torbjörn Alexandersson, F.M., who was a perfect co-diver during the investigations in 1960-62, and my brother-in-law, Mr. Gunnar Jansson, who accompanied me on the field work each summer. In his capacity as a physicist and engineer he was most handy with the instrumental equipment and the boat's engine, and in another of his capacities—that of chief steward and cook—he was excellent, which meant much during our long voyages.

The soil analyses were carried out at the Geomorphological Laboratory with the assistance of Mr. H. Gauja, F.K., and Mr. J. Bauska. At the same laboratory fittings for the equipment on board were built by Mr. E. Zetterström. Chemical analyses were made at the Geochemical Laboratory of the Swedish Geological Survey in Stockholm by Dr. S. Landergren and Dr. F. Manheim. Statistical data were compiled and bottom profiles constructed from echograms at the Institute of Geography by Mr. V. Silins. The finished copies of maps and diagrams were drawn by the cartographer of the Institute, Capt. W. Tiit, and Mr. E. Ludvigsen, Mrs. H. Mattson and Miss K. Olsson. The block diagrams were re-drawn from my pencil sketches by Mr. S. Liljefors. The photographic reproductions were made by Mr. L. Åberg. The manuscript was typewritten by Mrs. R. Brodin, and the language of the thesis was revised by Mr. N. Tomkinson.

Diving equipment was placed at my disposal by the Royal Swedish Navy, which trained me in the Scuba diving technique, when I was in service with the Naval Reserve. Financial contributions for the field work were received from the Andrée Fund of the Swedish Geographical Society, the Field Research Fund of the Faculty of Mathematics and Natural Science, and the Längman Fund which also has contributed to the printing of the plates. The cost of revising the language of the thesis was born by the Uppsala University Fund for the Promotion of Postgraduate Work, and for its publication a contribution was received from the State Council for Scientific Research.

METHODS

General

The principal subjects discussed in this paper are primarily based on field studies. For the general description of the Vättern basin information from various sources was compiled; ordnance-survey, land-survey, economic and geological maps, charts and air photographs, meteorological and hydrological statistics, papers dealing with the topography, geology and physical geography of the basin. Observations were also made in the field, especially with regard to the general morphology and the geology.

The field studies on the shore and bottom morphology were mainly carried out by commonly used methods. A large part of the studies of nearshore morphology in shallow-water areas was carried out by Scuba diving technique. This technique is also commonly used, but it was not as common as it is to-day, when the bottom investigations started in 1956, and its applications to subaquatic research is as yet not standardized.

The field methods had to be accommodated to the equipment available, and to the fact that the investigations were carried out by one man. As has already been mentioned in the introduction, the equipment during the first few years was quite inadequate, and the assistance of interested friends and students could only be obtained during the summers.

The field studies may be characterized as geomorphological mapping and sediment sampling. The sediments were chiefly analysed with respect to grain-size distribution, because this characteristic of the sediments can be regarded as the one which principally links the morphology to the processes by which it has been formed. In the following section the field methods and the sediment analyses will be discussed in brief.

Field methods

Shore surveys

Shore profiles were levelled with a Wild N10 and a Zeiss Ni2 level. The Wild level, having an azimuth scale, was also used for mapping. From 1960 a Wild RDS reduction tacheometer theodolite was used for this purpose. This instrument is very handy for investigations in cliffs, because distances reduced to the horizon are immediately obtained for vertical angles less than ± 45°.

When possible, the height of the profiles was levelled with reference to height stations of the Geodetical Survey, which give the height above sea-level to 1/100 of a metre. Levellings down to the shore were made over distances of 3 km. The height differences of the control levellings did not exceed 3 cm. However, in some places it was not possible to use the height stations. On Visingsö island, for example, there is no station. The height was then referred to the water level. Records of the water stage were obtained from the Motala power station. Because of the seiches of the lake the water level is not a fixed reference surface. The short periodic variations were estimated by a stick placed near the shoreline. The maximum variation observed during a profiling survey was about 10 cm.

By using an extension rod joined to a measuring lath and by having the man holding it dressed in a diver's suit, it was possible to level shore profiles to a depth of 1.4 m. By this method the depth interval, which is difficult to echo-sound, can be measured.

Most of the shore sediments were examined and classified by ocular estimation in the field. Samples of cobbles and coarser material were also examined in the field with regard to petrographic composition, individual weight and morphometric characteristics. Petrographically, arkoses, sandstones, pelites, shales and crystalline rocks were distinguished. Morphometrically, three-axis measurements were made.

The shore stratification was studied in shafts. It was generally impossible to work below the ground-water table, which was, as a rule, found slightly above the lake level. One remarkable exception was the clayey moraine of Visingsö. For example, in a shaft lying 1 m from the shoreline it was possible to go down 1.3 m below the lake level without having the shaft too rapidly filled with water. This shows the very low permeability of this material.

The amount of material in the range from granules to cobbles of moraine deposits was determined by wet sieving in the field of samples of 5 or 10 kg. The boulder content was estimated from vertical cliff walls.

Bottom surveys

Preliminary surveys

The following methods have been used in the preliminary surveys in order to obtain a general synopsis of the distribution of different bottom conditions.

1. Soundings with a recording echo-sounder
2. Scuba divings.
3. Free swimming in a light diving suit.

Navigation with continuous echo-sounding makes it possible to survey large areas quite quickly. The Swedish Hydrographic Office has found it possible to use as a standard a speed of 15 knots in their mapping surveys, and good results have even been obtained at a speed of 20 knots (oral communication from Mr. Fagerholm of the Hydrographic Office). In Lake Vättern the speed during preliminary surveys never exceeded 8 knots. The method was used in areas with a depth greater than 2 m.

The configuration of the profiles sounded facilitates morphological interpretation, and from the character of the echo different types of bottom sediments can be distinguished (cf. Hough 1952). In Lake Vättern it has generally been possible to distinguish the following types.

A. *Soft bottoms*
   1. Gyttja
   2. Recent fine inorganic sediments (Glacial varved clay)
   3. Submerged peat

B. *Hard bottoms*
   1. Sand (Glacial varved clay)
   2. Stony bottoms and moraine
   3. Exposed bedrock

The echo from the soft bottoms is broad, and has a somewhat diffuse contour, especially when the bottom consists of gyttja. Often an echo of a lower discontinuity surface of the sediment is also recorded (cf. Fig. 133). Submerged peat lying on the bottom has been found to give a somewhat diffuse echo contour, but the echo width is less than that of the other soft bottoms. How the echo from peat varies with the composition and thickness of the peat deposits is unknown.

The sand bottoms give a distinct echo with an even contour. The echoes from Glacial varved clay have been found to vary considerably. Sometimes they appeared to come from a real soft bottom, sometimes they looked like an echo from a sand surface. This may reflect sedimentological variations, but can partly be attributed to the adjustment of the sounding apparatus and the use of different types of recording papers.

Like sand, stony bottoms give sharp echoes, but the contour is irregular in the details. This type of bottom is often impossible to distinguish from a moraine surface. This is natural, because most of the stony bottoms consist of a residual layer of stones resting on moraine. The exposed bedrock gives echoes which are very much like the echoes from sand surfaces but are generally easily distinguished from these by the configuration of the profile.

The constituents of the bottom cannot be reliably determined only by the character of the echo, but, together with the knowledge of the depth and the distance to the shores, and in comparison with bottom samples, the interpretation of echo-sounding diagrams makes possible a general synopsis, on which further investigation plans can be based. It is comparable with air-photograph interpretation in subaerial investigations.

In areas of submergence morphological relicts of former shores can be expected to be traced by echo-sounding. It has been possible to identify fossil subaquatic bars and abrasional terraces.

Since the second world war Scuba diving operations (Scuba = self-contained underwater breathing apparatus) have more and more become a standard technique in investigations of shallow-water areas (Parker 1962). The foremost advantage of this technique is the possibility of making direct ocular examinations without disturbing the bottom. The old heavy diving equipments made movements on the bottom very difficult, and at every step a cloud of sediment particles was whirled up. With the light Scuba equipment, the diver swims freely, and thus he can cover much larger areas. The previous inability to examine the bottom directly was probably the most marked difference between morphological investigations above and below the surface. Being able to observe the ground is such a primary condition, that, from a subaerial point of view, the mere ocular observation of the bottom does not seem much of a technical advance. However, when compared with the possibilities of describing the bottom, when only scattered bottom samples can be taken from the surface, this diving technique means a large step forward. There have been papers which speak of disappointment with the results obtained by Scuba diving operations (Kidson, Steers and Flemming 1962). It should be remembered that diving, as such, is not a scientific method, and is no vademecum either, but, when used in the right environments for the right purposes, it can give valuable information which cannot be obtained from the surface. It should also be stressed that the diver must be the scientist himself or a person trained for special investigations by a diving scientist.

Diving operations can only be carried out to limited depths, but generally it is in the quite shallow areas that the rich morphological differentiation is found and also the most sudden transitions of forms and sediment composition. At great depths echo-sounding, bottom sampling from the surface, and photography with submerged cameras and flash-light are generally sufficient.

The maximum depth of diving is determined by the type of breathing apparatus. With the type which has been used in these investigations (AGA Divator), working with compressed air, the maximum depth is about 50 m, but normally work should not be done below 30 m. The practical working depth also depends on the natural light conditions and visibility, which is reduced by suspended material. The use of portable searchlights has proved to be of small value; when suspended material is present, the light has a purely negative effect because of reflection. In surf observation may be impossible because of the suspended material. In the Vättern investigations the working conditions were good down to about 25 m. The greatest depth at which observations were made was 38 m.

In the type of diving apparatus used the diver has an air supply of 1500 litres. The air consumption increases with the depth, varies with the type of work, and is also very individual. Even light work for which concentration is needed, for example, the taking of pictures, increases the air consumption. Normally about 75 litres/min are used, which gives a duration of about 20 min. According to an American collation of 3 000 dives, the average period under water was 15 min, the average depth being 17 m (Dill 1956, p. 279). In order to get a longer period of action, the diver can be supplied with air through a hose, coupled to the breathing apparatus. We used a 60 m hose. The area of action is naturally limited, when the supply is given from an anchored boat, but the hose is of little hindrance to the diver. For detailed examinations the use of a hose is generally to be preferred.

When swimming over the bottom, the diver has only a small area in sight, partly because of the range of visibility in the water, and partly because it is rather difficult to keep the same level, when swimming far above the bottom. Therefore the ability to orientate is limited, and it is difficult for the diver to estimate the relative positions of points lying more than a few metres apart. When it is rather dark, the visibility poor and the bottom diffuse (muddy bottoms), the consciousness of gravity becomes uncomfortably weak, and the ability to make exact observations may decrease.

Observations can be recapitulated in a report immediately after the operation or by telephone reports during the dive. The diver can also make sketches on a board during the operation, make notes about measurements, etc.

In the Vättern investigations a telephone connection was only used for a short time in the summer of 1962. The apparatus has a two-way connection, the diver's microphone being attached to the crown of the rubber hood, by which the noise from outblown air is minimized. On board there was a loudspeaker.

For a rapid estimate of the general bottom conditions towing the diver behind a boat can be very useful. I successfully tried this in the Baltic in 1955 down to depths between 15 and 20 m. The method has also been used in rivers with a rather high flow velocity (cf. Arnborg 1957). The diver can hang on to a diving plane (a simple piece of plywood or plastic with handles), which he uses as a deep rudder, or he can lie on a diving sledge. The towing speed should not exceed 3 knots (1.5 m/sec) because of the drag forces on the diver and his small range of visibility. In the summer of 1962 experiments were made with a surveying unit consisting of a towing boat with an echo-sounder and a diving sledge with telephone connection to the boat.

The operator at the echogram is in contact with the diver, whom he can inform about the bottom contour in front of the sledge, and the diver can give information, which can be directly noted on the echogram. The sledge should carry self-anchoring buoys, which can be released by the diver. The results of the experiments are encouraging, but improvements should be made on the sledge.

The swimming diver can document his finds by photography, by measuring forms, and by bottom sampling. The following types of cameras have been used: an Edixa Reflex (24 × 36 mm) with a wide-angle lens, housed in a specially made case of acryl plastic with a plane glass window (there is also space for an exposure meter), a stereo camera of the Stereorealist type (24 × 24 mm) in a case of metal and acryl plastic (Arnborg 1957, Fig. 1), and finally the commercial Rollei-Marin (6 × 6 cm). The cameras are equipped with flash-bulb units. The first-mentioned camera has proved to be the one easiest to handle. As a rule positive daylight colour film was used. The pictures were taken in natural light or with the use of white flash bulbs. The colour film gives better results than black and white film, especially when the bottom is poor in contrasts.

In the first few years of the Vättern investigations, when I only had a rowing boat at my disposal, a large part of the preliminary surveying near the shore was made by free swimming in a dry diving suit and with knife and camera as the only equipment. By blowing up the suit it was possible to swim on the surface for hours and make observations of the bottom down to at best a depth of about 10 m. For a closer examination of the bottom sediments and for photographic shots the suit was emptied of air before diving. Dives could only be made with a good result down to a depth of about 5 m. This technique is, of course, very primitive, and it only permits very simple observations, but it is useful for single-person surveys, and the equipment can be easily carried to remote shores.

Profiling by echo-sounding

A supersonic echo-sounder (Atlas Marin Monograph 658 Tr spec.), working on 50 MHz, was used. It has a frequency indicator and a hand-operated modulator. The depth can alternatively be recorded in the scales 1: 200 and 1: 400. Correctly adjusted, it has proved to be able to record the depth within ± 0.1 m down to a depth of 50 m. The transducer was placed in a special permanent rig at the side of the boat, and at 0.3 m below the still-water level. The Hydrographic Office has also found this position to be the best. There are more disturbing air bubbles under the bottom of a small boat than at its side.

From the transducer the supersonic waves are emitted in a cone of 30°—40° directed towards the bottom, and the least distance between the transducer and the bottom within this cone is recorded as the depth. Thus the recorded depth depends on the relation between the width of the cone, the true depth and the configuration of the bottom. The recorded profile may be deformed in relation to the true profile. The edge of a plateau, which ends with a steep slope will be laterally displaced towards the deep area. These deformations are generally of minor importance, but the angle of very steep slopes cannot be determined from the echograms. It is more important that narrow cavities cannot be recorded, because the cone covers the area around the cavity. The greater the depth, the wider the cavity must be to be recorded.

The length scale of the echogram is a function of the paper-feeding velocity and the speed of the boat in relation to the bottom. A postulate for a uniform scale is a constant speed in relation to the bottom. If it is assumed that constant r.p.m. of the boat's engine give a constant speed through the water, the variations of speed in relation to the bottom will depend on the wind, wave and current variations over the distance travelled. The problem of uniform length scale has been solved in the Atlas Ekolog 672, in which the speed of paper feeding is controlled by a unit with a transmitter and receiver on board and ashore, which continuously feeds the change of distance to a servo-motor. This combined echo-sounder and log of true motion in relation to the bottom has proved to be very valuable in morphological work (Sundborg and Norrman 1963), but unfortunately it could not be afforded for the Vättern investigations.

With an ordinary echo-sounder the length scale in relation to the speed through the water can be calibrated to the revolutions per minute (r.p.m.) of the boat's engine or, better still, the scale can be determined by the use of a log. In the Vättern investigations a precision propeller-type current meter (Ott Kempten "Arkansas") was adapted as a log. It was mounted on a steel tube 0.5 m in front of the boat and 0.3 m below the surface. The revolutions of the propeller are electrically recorded by a counter. The current meter was calibrated for two fixed engine speeds. In a distance of 300 m the maximum error was found to be 0.5 %.

When there is a uniform or variable current, the distance from the log will be erroneous. Then the echo-sounding should be done along a series of buoys with positions determined from the shore with a theodolite. The passages of the buoys are marked in the echogram (by short-circuiting the pen circuit), and the time between the passages is determined with a stop watch. The scale can be assumed to be constant between every single pair of buoys, or a smoothed correction curve can be calculated.

In profiling the boat can be steered by the compass or towards a fixed point ashore. If there is a constant drift, the boat will keep a straight course inclined to the head direction with compass steering, but with steering towards a fixed point the course will have a looped form. If the boat is steered by artificial marks in line with each other or along a straight row of buoys, the speed in relation to the bottom will be constant, if the drift is constant.

As far as possible the steering was done by natural or artificial marks in line with each other. When short profiles, transverse to the shores, were sounded, buoys were used as a rule. The soundings were made in good weather. The positions of the sections and the scale were determined by the following methods (with decreasing accuracy).

1. Course between buoys, with positions determined from the shore, and logging.
2. Course between buoys of known position.
3. Course towards marks in line with each other and logging.
4. Course towards a fixed point or a buoy with speed calibrated in r.p.m.
5. Course between fixed points.

When profiles transverse to the shore were echo-sounded in connection with bottom sampling, the following method was ordinarily used. The boat is steered towards the planned section with the echo-sounder in action. From the echogram the sampling positions can be determined. At the point of the profile nearest to the shore a buoy is placed. It must be possible to pass this buoy with constant speed without risk of running ashore. If it is a long profile, marks in line with each other are put up on the shore. The distance from the shore to the buoy is measured or estimated. The boat is navigated in the planned section with the echo-sounder in action, and buoys are placed where samples are to be taken, and at the outer end of the profile. The profile along the buoys is now sounded and the position of the buoys marked in the echogram. The positions of the buoys are determined from the shore, or the length scale is taken from the log figures. Then samples can be taken at the buoys.

Bottom sampling

Sampling from the surface

There were two hand winches on board, one small with 25 m of wire, and one large with alternatively 120 and 80 m of wire. The large one was placed on a pivot davit.

Samples of the bottom surface were taken with a modified Petterson dredge fitted with an extra weight to stabilize the dredge when lowered. The maximum sample thickness which can be obtained with this grab sampler is about 10 cm. In sand it takes surprisingly undisturbed samples. When the bottom material is badly sorted and contains small pebbles, it often does not close well, and the fine-grained part of the sample is lost. Hard clays cannot be taken.

Cores of soft, cohesive, inorganic sediments and of gyttja have been taken by a piston core-sampler designed by Professor Kullenberg and built at the Oceanographic Institute in Göteborg. The inner tube of clear, polished, acryl plastic has a length of 100 cm and a diameter of 5.8 cm. When taken out, the tube is closed with rubber caps. The sampler is released 2 m above the bottom by a forerunner weight, and falls freely into the bottom. The sample is held in the tube by the suppression caused by the piston, and there is also a spring shutter in the mouthpiece. It was found impossible to use the sampler, when there was more than 10–20 cm of medium sand or coarser fractions at the top of the bottom sediments.

Thus, from the surface it was possible to take sandy samples with a thickness of about 10 cm with the dredge, and samples of fine soft sediments with a thickness of at most 1.2 m (the mouth-piece included). Very compact fine sediments, poorly sorted coarse sand, and really coarse material were impossible to sample from the surface. However, all these are generally found in the shallow areas, where samples can be taken by a diver. The only type of material which was really problematic was fine sand. A 10-cm sample can be obtained by the dredge, but it would have been most valuable to get cores of these sediments. However, attempts with the piston coresampler were not very successful.

Sampling by diving operations

Diving technique makes possible a direct estimation of the differentiation of the bottom sediments. The diver can examine large areas and then estimate where significant samples should be taken. Less samples are needed when they are taken directly on the bottom than when taken from the surface. For example, if the bottom mainly consists of a residual layer of gravel and stones on a clayey moraine, but there also are patches of drifting sand, the diver can find out the average thickness of the residual layer, estimate the part covered by sand, and then take the samples, which are typical of this type of bottom.

In the Vättern investigations samples were taken in plastic bags with a volume of 1 litre, and in open brass tubes with an inner diameter of 6.0 cm and a length of 17.0 cm. When collecting sand and silt in the plastic bags, the upper 5 cm of the sediment was taken. In the bag any original stratification is, of course, destroyed. Hard cohesive sediments and peat were cut with a diver's knife. The brass tubes were used to get undisturbed samples. The tube was forced into the bottom with a screwing motion, and then a rubber cap was placed over the top of the tube. This was dug out and another cap placed over the lower end. The top of the tube was marked by its number. Samples were taken in this manner down to a depth of 22 m.

Because the diver is weightless in water, he cannot force a tube much further than 17 cm into a sand bottom. It could possibly be done with a pneumatic hammer or a vibrator. Simple digging was done in a search for discontinuity surfaces, but it is difficult to keep the shafts "clean". An air injector, supplied with air from the surface by a hose, was used in this work.

Here some simple measurements made by the diver should be mentioned. Dip and strike were measured with an aluminium bar and a compass equipped with a clinometer and a centimetre scale. This equipment was also used for the measurement of the length, height and orientation of ripple marks.

Detailed morphological mapping in shallow water

Generally a bottom survey can be carried out with sufficient accuracy by using a recording echo-sounder (Saville and Caldwell 1953). However, a bottom which is irregular in details cannot be morphologically mapped by echo-sounding. Narrow, deep hollows are not recorded, and the characteristics of the forms cannot be properly identified.

For the adequate reproduction of a bottom with a morphology of narrow furrows, the following requirements for the mapping were established by a preliminary diving survey. It should be possible to map forms of dimensions larger than 0.2 m. The measuring error of individual details should be less than ± 0.05 m, and the positional error within the total mapping area should, as regards the horizontal plane, be less than 0.25 m and, as regards the vertical plane, less than 0.05 m. These requirements exclude echo-sounding and other ways of sounding from the surface, firstly because of the difficulty in getting correct horizontal coordinates, and, secondly, because of the impossibility of identifying the configuration of individual forms from the surface. In order to overcome these problems a method based on direct measurements on the bottom and lead-sounding was used. The method requires diving equipment, but technically it is not complicated. It will be described in the chronological order of the different phases.

1. Along two sides of the mapping area rows of buoys are placed at about the same distance in each line. If the lines are transverse to the shore, marks in line with each other should be placed on the shore to give the direction of the rows.

2. The anchor lines are straightened by a diver, who also controls the stable position of the anchor weights.

3. The exact position of the buoys is determined by triangulation from the shore.

4. Between each pair of buoys on the opposite sides thin wires are stretched along the bottom. The wires have a mark for every half-metre. They are fastened to rods driven into the bottom at the anchor weights. Thus the wires form a set of parallel sections.

5. Parallel lines are drawn on white-painted

masonite boards, and the lines are divided by half-metre marks on a suitable scale.

6. The diver is equipped with the board, a flexible steel measuring tape, a compass with clinometer, and a pencil. He follows a stretched wire along the bottom, and draws the profile directly on the board. The lengths are obtained from the marks on the wire and the measuring tape, and the heights are measured in relation to the wire. Slopes can be measured with the clinometer and are noted on the board, as well as other observations of special interest.

7. The levels in relation to the water surface are determined by lead-sounding from a dinghy. The lead hangs in a hand winch with a depth indicator. The lead must be so heavy that the sounding line can be drawn to a vertical position with the lead resting on the bottom.

The lead is submerged to about 0.5 m above the bottom. Then the diver brings it (and the dinghy) to positions along the wire, where this is in contact with the bottom. By a signal from the diver the lead is lowered to the bottom, the depth is read in the boat, and the diver notes the position on the board. The number of sounding points depends on the relief of the section.

8. From the profile drawn by the diver and the sounded points, a profile is drawn and corrected with regard to levels in relation to the water surface. If necessary, the length is corrected in relation to the distance between the buoys. It may also be necessary to place some buoys in the section.

9. The corrected sections are drawn below each other on a board, and the diver goes down and connects forms between the sections. He may find that it is necessary to densify the system of sections.

10. A map can now be constructed from the parallel profiles and the observations the diver has noted about the morphology between the sections.

This method requires a diver who is familiar with morphological forms. The work takes such a time that the diver should be supplied with air from a boat. The diving operations can be carried out when there is some wave motion. The waves mainly disturb the diver by generating turbid water near the shore, which reduces the visibility. Really calm weather is only necessary during the sounding. The principal disadvantage of the method is undoubtedly that it is laborious. It would be expensive to use it for mapping of large areas. The field work for the profiles and the map in Plates 2 and 3 was done by three persons in eight effective days, which means that in reality it took 15 days.

Sediment analyses

Sampled sediments were principally analysed with regard to their grain-size distributions. This distribution was determined by sieving and by pipette analysis.

Swedish standard sieves with square meshes of 16, 8, 4, . . . 0.0615 mm, equivalent to average particle diameters of 20, 10, 5, . . . 0.078 mm, were used. When the sediment was obviously free of organic material and there was an insignificant amount of particles finer than sand, the samples were dried at 105°C and then directly sieved. Samples that held fine particles or organic material were dried at 105°C, weighed, treated with hydrogen peroxide, dried and weighed again and then sieved.

From sieved samples 5 g of the material that had passed the 0.0615 mm sieve were taken for pipette analysis. The material was dispersed in the pipette cylinder. To the distilled water sodium oxalate was added as peptisator. The cylinder was turned in a machine for 10 hours. The pipetting was then done in a constant-temperature room (20°C) for every Ø unit ($Ø = -\log_2$ diameter in mm) from 4 to 10 (0.0625 to 0.00049 mm) or to the 95th cumulative percentile, if the amount of clay was insignificant.

Samples with a large proportion of clay were washed with distilled water on a membrane filter no. 7 (0.2 μ). In no case were any particles visible in the washing water that had passed the filter. This treatment was used because some samples did not settle without it. Their electrical resistance did not prove any content of chlorine, so the reason is unknown.

The content of organic material was in some cases determined by weighing before and after heating to 400°C.

The special analyses of a single core of varved clay are dealt with in the chapter on the glacifluvial sediments of the northern and central parts of the lake.

The cumulative grain-size distributions were drawn on probability paper with the Ø units in a linear scale. The statistical characteristics used in this paper are also based on the Ø units (Inman 1952):

$Md_\varnothing$ = Median (50th percentile on cumulative curve)

$M_\varnothing$ = Mean $= 0.5\ (\varnothing_{84} + \varnothing_{16})$ (84th and 16th percentiles)

$\sigma_\varnothing$ = Deviation measure (sorting) $= 0.5(\varnothing_{84} - \varnothing_{16})$

$\alpha_\varnothing$ = Skewness $= (M_\varnothing - Md_\varnothing)\sigma_\varnothing^{-1}$

The statistical grain-size characteristics of samples referred to in the text and will be figures found in Table 26 at the end of the paper.

THE VÄTTERN BASIN

Geology and geomorphology

Introduction

The geology of the Vättern basin is best known from the maps with descriptions edited by the Geological Survey of Sweden in the Aa Series, on a scale of 1: 50 000, but there have also been several special investigations. The geological maps are not very up-to-date; nine of the eleven sheets covering the area were mapped before 1907, one in the 1920s and one in the 1950s.

There are many unsolved problems as regards the structural geology and the development of the basin, but it is beyond the scope of this paper to investigate the geology in detail, especially that of the bedrock. However, the rocks are of such importance to the geomorphology of the shores that a general description of the area is necessary. This description is mainly based on the geological maps (not referred to in every single case), on the special investigations dealing with the area, and on the author's own observations. The crystalline rocks and the sedimentary rocks are described separately, because their morphology differs significantly, and because the effect of the shore processes on these categories is totally different.

The Quaternary geology is partly the subject of this study. All the soils in the basin are of Glacial or Late-Glacial origin, and thus their distribution determines the distribution of the main sources of the shore processes. The bottom sediments of the present Lake Vättern cannot be explained without taking into consideration the Late- and Post-Glacial developments, as interpreted from earlier investigations around the lake, but, on the other hand, they can also add to our knowledge of these developments. The land uplift, which has been going on since the ice recession, is the process which largely determines the present shore features. The relation between the history of the Baltic and the Vättern basin and the development of the isolated lake stages are therefore dealt with in the special chapter on the Quaternary geology.

General topography

The Vättern basin is situated in southern central Sweden between latitudes N. 58°55′ and N. 57°45′. It has a pronounced elongation with the main axis directed NNE.–SSW. The width of the basin, the border land included, is only of the order of 30 to 40 km. The basin and the surrounding ridges separate the central Swedish lowlands into an easterly and a westerly part. The present mean water level of Lake Vättern is 88.5 m above sea-level, and the outlet is at Motala on the eastern shore.

Along the western shore up to Karlsborg runs an elongated hilly ridge, named Hökensås, which has a maximum height of about 300 m above sea-level. North of Karlsborg the border consists of a lower, dissected landscape (cf. Pl. 1). In this area there are different massifs. The largest and most dissected is that of the Tiveden Forest along the northern shore.

To the south the basin ends in two valleys separated by a hill. The highest bordering area is to be found along the south-eastern shore, where the hills reach about 350 m above sea-level. From this area the height gradually decreases towards the north, and the level Ostrogothian Plain from around the shallow Lake Tåkern and up to the surroundings of Motala lies at about 100–110 m above sea-level. In this area only "Mount" Omberg with a height of 263 m stands out. North of Motala is the dissected massif of the Tylö Forest.

Besides the Ostrogothian Plain, flat lowlands are found within the basin north-west of Motala, along the shore at Gränna, around Jönköping at the southern end of the lake, on the island of Visingsö and below Hökensås from south of Hjo up to Karlsborg. This last-mentioned area is called Guldkroken.

The island of Visingsö is the only island in

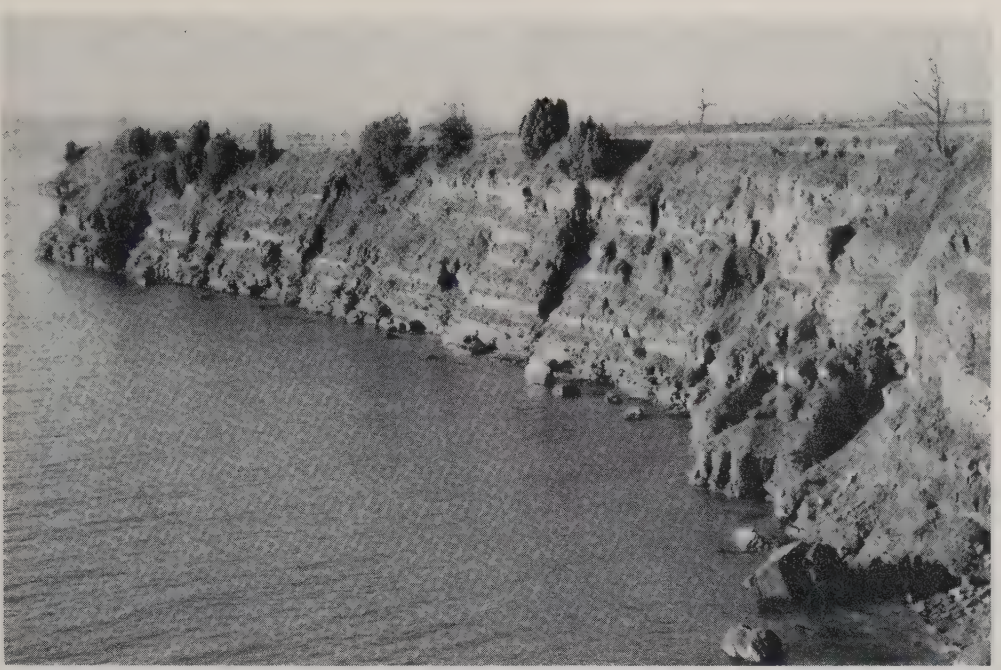


Fig. 1. The sandstone cliff of south-eastern Visingsö. April 1959.

the southern part of Lake Vättern. At the mouth of Motala bay is a series of small islands and skerries, and in the northernmost part of the lake a dense archipelago is found.

With respect to the bottom topography the basin can be divided into three sections: *the southern and central section*, south of Karlsborg and the northern foot of Mount Omberg, *the northern section*, north of the former and including the islands at the mouth of Motala bay, and finally *the Motala bay* (cf. Pl. 1).

The southern and central basin is characterized topographically by a deep channel and a plateau. The channel runs along the eastern shore and has a maximum depth of about 100–130 m. A short branch runs from the south to west of the island of Visingsö. The plateau, which is connected to the western mainland, has over vast areas depths between 30 and 40 m.

The northern basin is characterized by deep troughs and channels separated by ridges and platforms, and the topography is more irregular than in the previous section. The Motala bay is quite shallow.

The different topographical characters of these three sections reflect differences in their geology, and accordingly a separate description of the regional geology and geomorphology is given for each of the sections.

The rocks and general structural geology

The Vättern basin lies within an Archaean boundary zone between the gneisses of south-western Sweden and the volcanics and granites of south-eastern Sweden. The tectonics of this zone are characterized by eastward-directed overthrusts, probably of Jotnian age. The thrust planes strike parallel to the longitudinal axis of the lake and are represented by zones of schistose granite and gneiss, running along the present hillsides.

Directly on the Archaean crystalline bedrock a sequence of sedimentary strata, called the Visingsö Series (Nathorst 1879), is found. According to Collini (1951, p. 36), this Series has a total thickness of more than 1040 m, and can be divided into three main stages. From the bottom there is first a yellowish-white quartz sandstone (> 145 m), and above that a sequence of various grain sizes (> 315 m), with arkoses predominating but with intercalated pelitic layers. The colour varies from reddish-violet to green. Shales (> 580 m) form the highest part of the Series, their colours varying from grey to black.

In some areas the yellow quartz sandstone seems to be underlain by reddish sandstones and conglomerates (Brotzen 1941).

The shales contain scattered micro-fossils. Their age, as well as that of the whole Series, has been discussed ever since the first investigations (cf. Wiman 1894, Rosén 1927, Ewetz 1932, Martin 1939, Collini 1951). According to Timofeev (1960), the fossils are of Eocambrian origin, but their age cannot yet be considered to have been finally determined.

The present distribution of the Visingsö Series is limited to the basin, except for a few small finds in the neighbourhood. The distribution was fairly well described as early as 1819 by Hisinger. The sediments above the lake level appear as fringes between the shore and the hillsides. Rocks of this Series are also found on some islands in the northern section, and the island of Visingsö is entirely made up of them.

The whole Series seems to have been deposited in a continental basin, where the climate changed from arid and partly cold during the deposition of the sandstones to warm and humid during the deposition of the uppermost stage, when the clays were formed and even carbonate settled. Mud cracks and other evidences of subaerial influences show that the sedimentation, at least in part, took place under shallow water and nearshore conditions.

The present position of the sediments is partly due to large tectonic movements of gravity fault and rift character. The Vättern basin has often been taken as an example of a typical rift valley or graben, bordered by parallel fault scarps (S. De Geer 1910, Cloos 1936). However, according to the investigations by Martin (1939) and Collini (1951), the western border represents a denudated Jotnian nappe, and no traces of a rift with an effect on the Visingsö Series have been found there. But at the eastern margin a rift, caused by tension within the old thrust zone, can be followed along the full length of the basin with horses of Visingsö strata, forming benches between the exposed fault scarp and the lake. The maximum slip has been stated to be at least 1000 m (Collini 1951, p. 40). This single rift, documented from observations on land, does not exclude the possibility of a true rift valley within the area of the present lake surface.

The age of these tectonic movements is uncertain. According to Asklund (1931, p. 369), they are Late-Precambrian. Martin (1939, p. 52) suggests that they mainly occurred in the Permian, but again Collini (1951, p. 40) only mentions Late-Precambrian movements (when discussing the southern part of the basin).

The Ostrogothian Plain east of the Motala bay is built up of Cambrian, Ordovician, and Silurian marine sediments, which can also be presumed to cover the bottom of the bay. From a geological point of view, this plain and the bay can be regarded as not belonging to the Vättern basin proper (Nathorst 1879). The Cambrian-Silurian Series consists of sandstone, shales and limestone, and the Series entirely rests on Archaean rocks, as far as is known from the relatively few drillings (Thorslund 1951). Bearing in mind the nearness to the thick Visingsö sediments, this is rather surprising, and can only be explained by a complicated pattern of tectonical movements of different ages.

The Paleozoic sediments have been preserved below a steep fault along the northern margin of their present distribution. The faulting probably involved an upthrust of the northern block. This E.-W. fault belongs to a large Post-Silurian system with the same direction, which lends character to the central Swedish lowlands north-east of Lake Vättern (S. De Geer 1910, Asklund 1927, 1931).

The morphology of the isolated troughs of the northern basin must be regarded as having been influenced by a N.-S. pressure analogous to the forces that caused the E.-W. fault system. Traces of such a pressure are also found in the structure of the Visingsö Series, in the joints of the sandstone on the island of Visingsö, in the pattern of dips of the strata south of Gränna (Collini, op. cit., p. 40), and in the folds of the sandstone and the Archaean substratum at Huskvarna (Brotzen 1941). All these deformations are not necessarily synchronous with the large dislocations north-east of the Vättern basin. As already mentioned, Collini considers the movements in the southern part to be synchronous with the N.-E. Precambrian faulting.

Regional bedrock geology and geomorphology

The Archaean rocks of the southern and central section

Mount Omberg prominently marks the beginning of high cliffs along the eastern shore. This granitic hill is a horst, and in the steep walls facing the lake slip planes and brecciated rocks are exposed. The subaquatic slope runs down to 60–80 m and then flattens out to a maximum depth of about 100 m.

In the small bay south of the horst (the harbour of Hästholmen) the porphyritic rock is much dissected, morphologically reflected in skerries and an irregular bottom with vertical subaquatic walls, 30–40 m in height, off Hästholmen (M in Pl. 1).

From Hästholmen southwards a straight cliff is formed of the granite and partly porphyritic rocks of the fault zone. The structure is exposed along the shore and in road cuttings, running at the top of the cliffs. The cliff line is fairly parallel to the N.–S. strike. The dip of the slip planes varies from about 50° to 90°. The level of the bordering land gradually ascends to the south, and the cliff leaves the shore at the latitude of the northern end of the island of Visingsö.

The present morphology is largely the product of glacial wear and plucking, and from a genetical point of view the escarpment is not a proper shore cliff. Because of frequent joints, the plucking effect predominates in the cliff walls, and the cliff descends in irregular steps. Post-Glacial weathering and rockfalls have been of minor importance.

The rock behind the cliff line forms a landscape broken into details, in which the glacial wear is represented by a sheepback morphology, but where the relief is still low. This surface is a remnant of the Precambrian peneplain, underlying the sedimentary rocks of the Ostrogothian Plain (cf. Thorslund 1951, p. 128, and Rudberg 1954, Pl. 1).

South of Gränna the morphology of the border land is more hilly with summits more than 200 m above the lake level. A high cliff with vertical walls forms the shore 7 km south of the town, but only for a few kilometres, then again a strip of low land separates the cliff from the shore. Granite constitutes the rock north of Lake Landsjön (which fills a depression near the shore), and south of this lake there is gabbro and diorite, changing into schistose granite in the neighbourhood of Huskvarna (Gavelin 1907). In the area of gabbro and diorite, the structural features of the fault do not stamp the morphology so markedly as, for example, in the area north of Gränna.

As previously mentioned, the western shore of this section is bordered by the Hökensås hills. The predominant rock is granite, but a gneiss massif in the area south of Bankeryd is also of morphological importance (E in Pl. 1).

Generally the hillside has smooth forms, and the bedrock is exposed at fairly few localities near the shore. The only area with a high rock cliff is that of the gneiss, where (north of D in Pl. 1) a head of naked rock stands out along the shore. The cliff is partly steep, but the forms are rounded by glacial erosion.

There are scattered outcrops of granite along the shoreline in the area north of the promontory of Svedudden (F in Pl. 1) to about the latitude of the northern end of the island of Visingsö. Angular forms with rough surfaces through glacial plucking predominate, but there is also some dressed rock. North of this area, the hillside leaves the shore and gives place to a flat lowland, entirely covered with glacial deposits.

The Visingsö Series in the southern and central section

The largest area, which from its general morphology and scattered outcrops can be attributed to the Visingsö Series, is connected to the middle part of the western shore.

As previously mentioned, a flat lowland borders the hill of Hökensås from south of Hjo to Karlsborg. The bedrock is not exposed anywhere in this area (called Guldkroken), but drillings have proved the occurrence of sandstone (Linnarsson 1880, Brotzen 1941). Off Guldkroken there is an even bottom with depths less than about 45 m. All recorded echo-soundings from the offshore bottom show extremely smooth curves without any irregularities. The area includes the island of Visingsö and is restricted to the east by the channel, which runs off Mount Omberg between Visingsö and the eastern mainland and down to the southern end of the lake (cf. Pl. 1). The even bottom topography indicates a

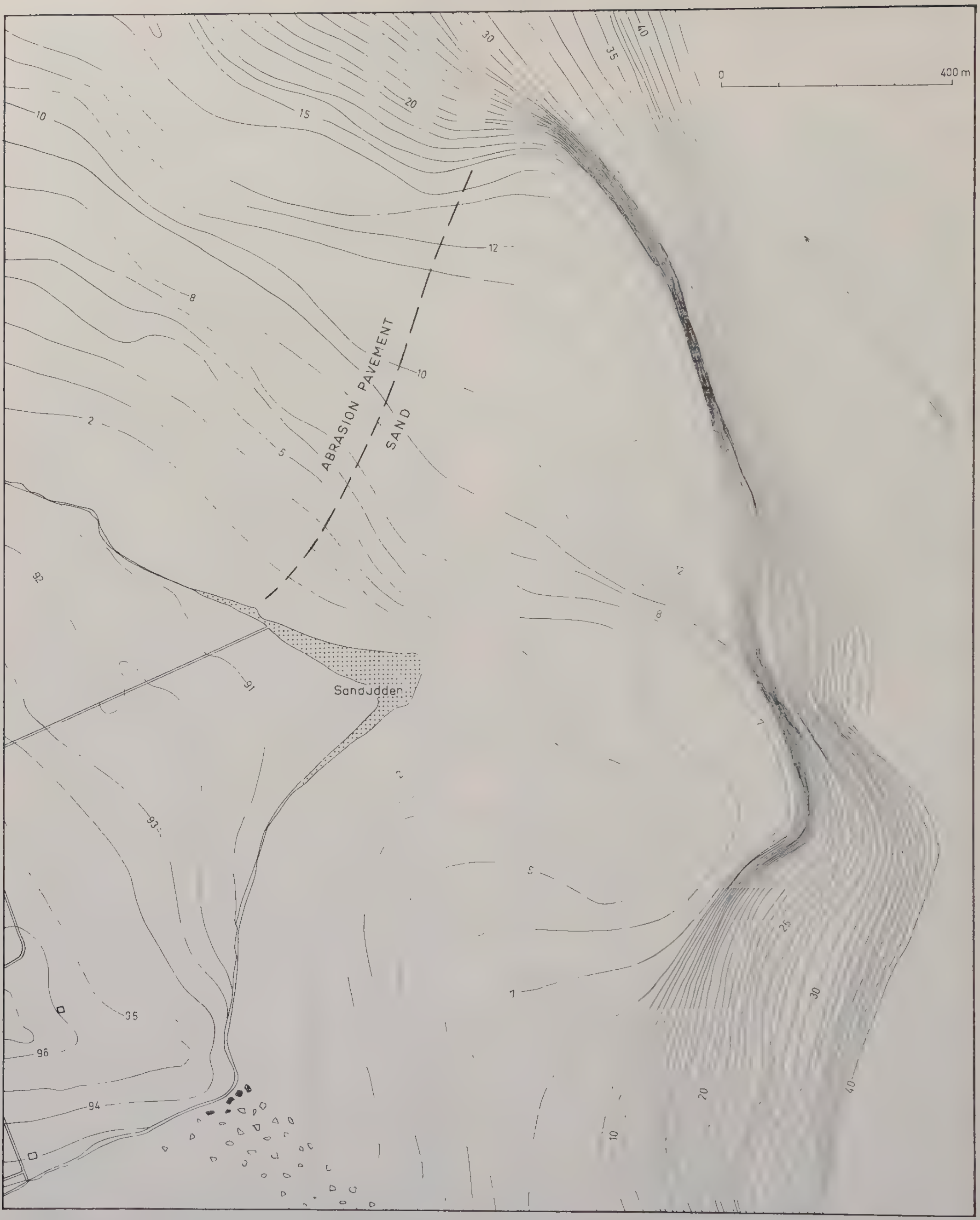

30
40
35
0
400 m
20
15
10
12
8
10
ABRASION PAVEMENT
SAND
2
5
12
8
92
91
Sandudden
7
93
5
25
7
30
35
96
94
20
10
40

Fig. 2. The area around the Sandudden ness on Visingsö island. Land-contour intervals in metres above sea-level. Depths in metres. From a map of the northern part of Visingsö island surveyed by the author (Norrman 1957).

coherent, slightly disturbed area of rocks of the Visingsö Series.

In 1961 at the Flisen shoal, halfway between Karlsborg and Hjo and 3–4 km from the shore (cf. Pl. 1), the exposed bedrock was examined by diving operations at depths between 1.5 and 4 m. The rock of this shoal is a fine-grained, violet-red sandstone with a main

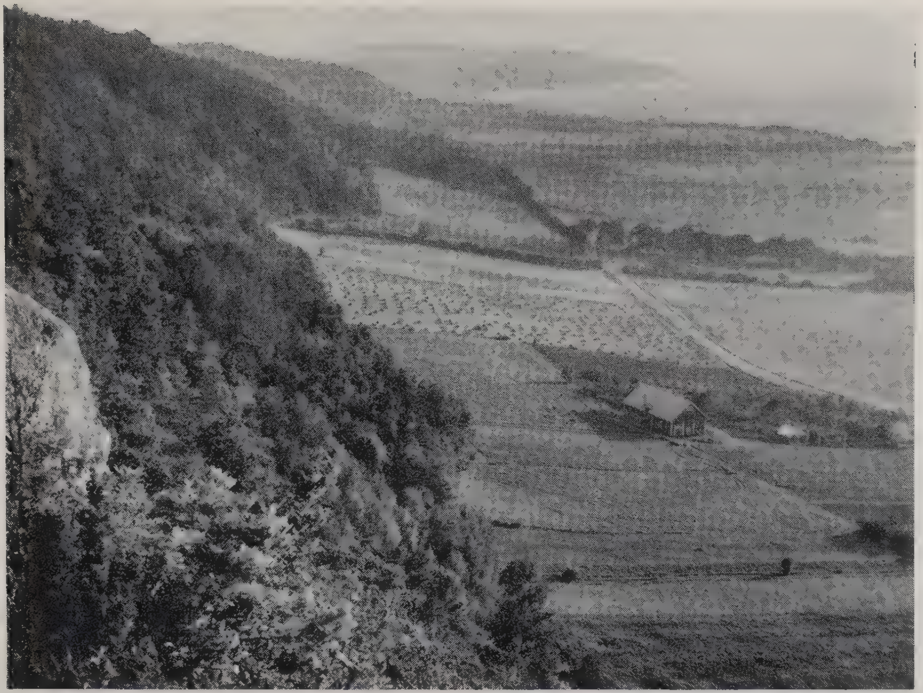


Fig. 3. View towards the south over the Gränna area. T. Alexandersson 1963.

strike of N. 20° E., and the beds dip about 5° from the shore. In comparison with samples from the island of Visingsö, this sandstone can be placed in the upper or central parts of the middle stage of the Visingsö Series.

Along the western shore of the island of Visingsö there are small outcrops of mostly ligth, yellowish sandstone at the shoreline. The dip is towards the north-east and is generally fairly steep (15°–30°). This sandstone is also exposed off the southern point of the island. Here the dip is less than 10°.

Along the south-eastern shore of the island, steep cliffs (Fig. 1) have been formed by abrasion of sandstone and pelite of the middle stage, northwards changing into shales of the uppermost stage. The dip is easterly with a steepness of 5°–10° (cf. Fig. 100). The bedrock is not exposed north of the Visingsö harbour.

North of the island, exposed beds of shale and limestone are found immediately off the shore. Outcrops have been examined down to 6 m in depth. The dip is easterly ($< 5°$–10°). At greater depths the rock is embedded in boulder clay and Post-Glacial deposits. The beds form a low ridge (called Getryggen) towards the NNE. along the eastern margin of the plateau.

The plateau ends very abruptly east and south of the island of Visingsö. Off the shallow bottom east of the ness of Sandudden (Fig. 2) the upper parts of the subaquatic slope are covered with sand, deposited at the angle of static repose. In lower, steeper parts (not examined), the rock is probably exposed. This escarpment runs almost straight along the full length of the island.

On the mainland east of Visingsö, the Series appears in the lowland below the fault scarp (Fig. 3). This platform, inclined towards the lake, hangs on the footwall with the strata dragged upwards along the fault plane (Martin 1939, p. 46). An excellent section from the granites at the main fault and down to the shoreline, cut by a rill 5 km north of Gränna, has been investigated by Collini (1951, 1958). In this section the transition from the lowest to the middle stage is exposed, and through faults within the Series shales of the highest stage have been brought level with the latter, near the present shore.

From 5 km north of this section down to

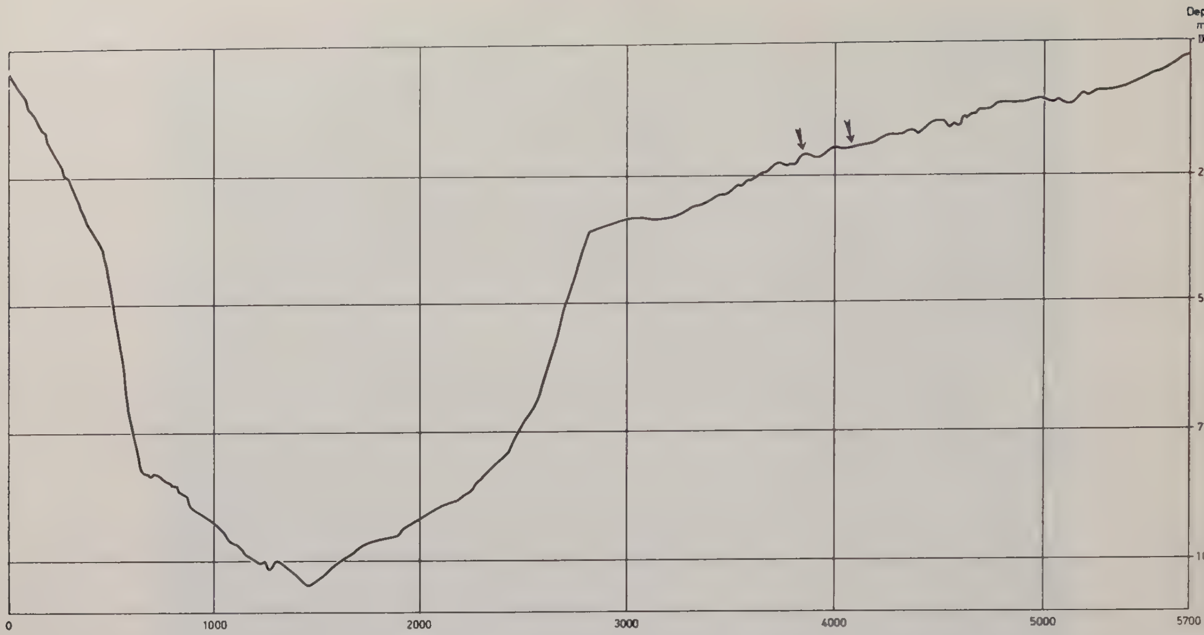


Fig. 4. Echo-sounded bottom profile between Visingsö island and Gränna harbour. The Visingsö shore is 200 m further to the west (left). The profile ends 1 450 m north of the entrance to Visingsö harbour. The arrows indicate the sites of submerged bars.

Gränna harbour, the Series is exposed on the shore. A cliff up to about 8 m in height has been formed in the shale, but the surface descends towards the south, and south of the harbour it is not exposed on the shore. The beds can be morphologically traced in a gently sloping subaquatic platform, which is broadest at the harbour with a width of 3000 m. The mean slope in this area is 1.17 %. As on Visingsö, the even surface ends abruptly (Fig. 4), and the channel between the western sedimentary plateau and the eastern fault zone probably reflects a real rift valley.

North of the Gränna area up to Mount Omberg, very few traces of the Visingsö Series are left. The cliffs are formed of the Archaean rock, and immediately off the shore, the depth is 30–90 m. At Mount Omberg, however, a narrow fringing horse of shale and sandstone is found dragged upwards along the steep wall, facing the lake.

South of the Gränna area the partly vertical granitic cliff descends into the lake, but there are some pockets of sandstone (Collini 1951, Fig. 8), and 14 km from Gränna harbour there is again a continuous narrow bench between the escarpment and the shore. This bench is only 100–200 m wide down to Huskvarna, where it widens to about 500 m. The bedrock, entirely hidden by till, consists of sandstone (Gavelin 1907, p. 86).

Sandstone can also be presumed to constitute the subsurface in the whole area of low land around the southern end of the lake. The sandstone is only exposed in some localities at the foot of the steep hillsides, but is also known from drillings made in water-supply investigations. The sandstone surface lies 54 m below the shore at the mouth of the Huskvarna River, and ascends in steps towards the north (Ullerstam 1940). A sequence of yellow and red sandstones, arkoses and conglomerates has been identified from the drillings. The structure of the sedimentary rock and the crystalline substratum of the Huskvarna area has also been reconstructed from these drillings, and, according to Brotzen (1941), the beds, as well as the crystalline subsurface, are deformed in E.–W. folds with the axis inclined towards the N.–S. fault.

In the western parts of Jönköping, the sandstone has been found at depths between 11 and

54 m and at the southern end of Lake Munksjön at 80 m; on the shore of Lake Vättern in the eastern part of the town a drilling was driven down 194 m before sandstone was met (Lindqvist 1940). This is the thickest deposit of loose soil known in Sweden.

At the western shore north of Jönköping, a strip of sandstone extends along the shore to Trånghalla (D in Pl. 1). These beds are reflected in the morphology of the nearshore sections (cf. Fig. 145). According to Martin (1939, p. 44), the yellowish-white quartz sandstone of this area belongs to the lowest stage of the Series, is of eolic origin, and rests on deeply weathered granite without any traces of displacement.

As has been proved by drillings, sandstone again appears in the valleys at Bankeryd and Habo (E in Pl. 1 and the first rivulet north of that), embedded in thick glacial and glaciofluvial deposits (Collini 1951, pp. 23–24). In the same position sandstone beds are presumed to constitute the bedrock of the flat bordering land up to the promontory of Svedudden (F in Pl. 1).

Off this area, north of Bankeryd, there is a shallow, even platform with a width of up to about 1000 m (cf. Fig. 71). Its level is partly conditioned by Quaternary deposits, but the entire formation seems to presuppose an even substratum, characteristic of the bordering, sedimentary rocks. To the north the platform is connected to the large subaquatic plateau.

The significant influence of the Visingsö Series on the morphology of the basin south of Karlsborg and Mount Omberg is obvious. Bordering lowlands, subaquatic platforms and plateaus, related to the presence of the sedimentary rocks, are everywhere a regular pattern. Taken as a whole, the level of these morphological elements is much the same within a few tens of metres. This level must be due to a general denudation within the basin. Nothing is known about the geology of the deeper parts, except for the drilling at the southern end of the deep channel in Jönköping.

Because of the relation to the thresholds of the basin, the morphology of the present channel must be regarded as a product of glacial excavation, which has most seriously affected the less resistant fault zone. With the present poor knowledge of the bedrock of the bottom, it is, however, impossible to state the proportion of loose material excavated and the relation between excavation within the channel and glacial abrasion of the "plateau level".

The Archaean rock of the northern section

The eastern and western borders are congruous with the borders of the southern part of the basin in their connection with the Jotnian thrust zones. In the northern section, the eastern zone has also been rejuvenated through faulting. The main trends in the bottom topography are also evident along the lake. However, there are also differences. The lowland and subaquatic plateaus along the western shore are missing, islands and ridges divide the bottom into separate channels and troughs, and finally there are the dissected strandlines and the archipelago.

Along the northern part of the eastern shore, the main rock is granite, but there is also gabbro, and in the northernmost part gneiss. The morphology is broken and hilly with narrow valleys, running mainly N.–S., eroded along fissures and local depressions. The islands (east and south-east of R in Pl. 1) have whaleback forms through glacial erosion. The subaquatic slope off these islands is steep, partly vertical (cf. Fig. 132) and formed within the fault zone. South of this area the escarpment leaves the shore, and a sharply defined hillside runs in a N.–S. direction east of a block of low land north of the Motala bay. The granitic ground of this area is very flat, except one single hill with a height of about 30 m.

The entrance to the Motala bay will be described in connection with the Visingsö Series.

The western or rather north-western border is dominated by the Tiveden massif, mainly formed of gneiss and granite. This massif is dissected by fissures and joints, running N.–S. and E.–W., which through erosion and glacial excavation give the morphology a rectangular shape. The main direction of the central and northern parts of the shore, running NE.–SW., is in contrast to this pattern. In the central part there are vertical cliffs along the shore, with a height of about 20 m. At the base of these cliffs, slip planes proving the Jotnian overthrust can be found (demonstrated by

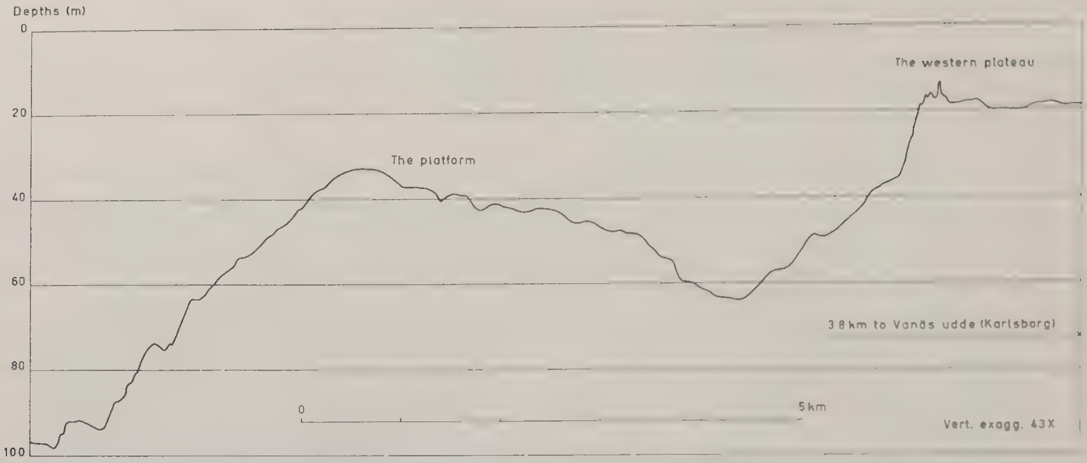


Fig. 5. Echo-sounded bottom profile in direction N. 34.5° E. from the light on Vanäs udde in Karlsborg.

Collini), but there are no structural traces of a gravity fault. The present wall must be regarded as the result of glacial plucking along vertical joints.

The subaquatic slope along this shore is steep, partly with vertical walls, interrupted by flat, narrow platforms in a step-like manner. The generalized shoreline and subaquatic contours in the central part and off the archipelago are straight, and this scarp is mostly regarded as the border of a downthrow (S. De Geer 1910, K. E. Bergsten 1943) but not by Rudberg (1960, Fig. 3.2).

This north-western shore with its narrow inlets and the swarm of islands and skerries is very much like the classical archipelago (skärgård) of a formerly glaciated, dissected landscape, found north-east of Stockholm. Viewed from the north, the islands appear as well-rounded bosses, but south-facing facettes have been roughened by dissective plucking. On these rocky island shores facing the deep troughs, the same joint system as at Tiveden is exposed, but also here evidence of a fault zone is lacking.

The southernmost part of the western shore of this section is of a different morphology. The area north of Karlsborg is a lowland with a morphology similar to that of the opposite block, north of the Motala bay. It is a granitic block of low relief, and the nearshore subaquatic platform has the same level as the plateau of the Visingsö Series south of Karlsborg.

The troughs and channels of the northern section have not been totally covered by echo-soundings, and the necessarily generalized contours of Pl. 1 do not show how broken the topography really is.

South of the northern archipelago is a deep, isolated trough.(A single lead sounding at its northern margin showed a depth of 114 m.) It is closed to the south by the island of Stora Röknen (P in Pl. 1), mainly formed of granite, the skerry of Lakaskär, a needle-shaped projection, and a subaquatic ridge east of Stora Röknen. This ridge separates the trough east of this island into two channels (cf. Fig. 132).

South-west of the island of Stora Röknen is another trough with a maximum depth of about 100 m. At its southern end it is separated into two channels by a subaquatic platform. The eastern channel runs in the direction of the large channel of the central section. The western channel follows the north-western shore and sends out a branch south-west of the platform. Echo-soundings across the platform (Fig. 5) have shown that the northern slope ascends in steps, the steps being inclined toward the south. The top of the platform also dips in the same direction. The slope of the plateau south of the platform is rather straight in the upper parts, but again there are steps in the lower part. The bottom of the channel is filled with muddy sediments. The section of Fig. 5 certainly gives the impression of a tilted fault block, which originally formed a part of the plateau. This presumed dislocation supports the hypothesis that the troughs of northern Lake Vättern are of tectonic origin, and have the character of small grabens, the faulting having been caused by tension transversely to older structural zones.

The Visingsö Series in the northern section and the rocks at the entrance to the Motala bay

Small outcrops of yellowish-white sandstone deposited on granitic rock are exposed on the northern end of the island of Stora Röknen and nearby islets. At the southern end of Stora Röknen this sandstone, which belongs to the lowest stage of the Series, has been quarried, and it is presumed to entirely cover the southern half of this island and the shallow offshore platform (Blomberg 1901). On the eastern mainland, the same sandstone is to be found along the western shore of the lowland north of the Motala bay, where it forms low hills (the hills are indicated by the closed 100 m contours in Pl. 1). The beds dip towards the SE.–SSE. with an inclination of 15°–20° (Jönsson 1887, p. 12). Here the sandstone is still quarried.

On the level ground south-east of the hills mentioned, sediments, classified as shales by Jönsson (op. cit.), have been found in small scattered outcrops. However, according to his description, they are rather identical with the fine-grained sandstone and the pelite of the middle stage.

South-west of these sediments, a conglomerate was found by Linnarsson (1880) on the shore and the nearshore bottom at the south-eastern end of the lowland. The stones of gneiss and granite are described as mostly rounded and "flattened out", lying in a matrix of weathered granite. The "shales", as well as the conglomerate, have the same strike and dip as the sandstone, which they thus cover. South-east and south of the sedimentary outcrops, granite is exposed on the mainland and the islets nearby.

The entrance to the Motala bay is marked by a ridge, where the island of Fjuk (L in Pl. 1) and the Jungfrun skerry (K) project above the lake level. The ridge has a rugged morphology with steep walls, as can be seen from Fig. 6. North of the Jungfrun skerry a gap leads to a channel running towards the outlet at Motala.

The rock of these islands was identified as

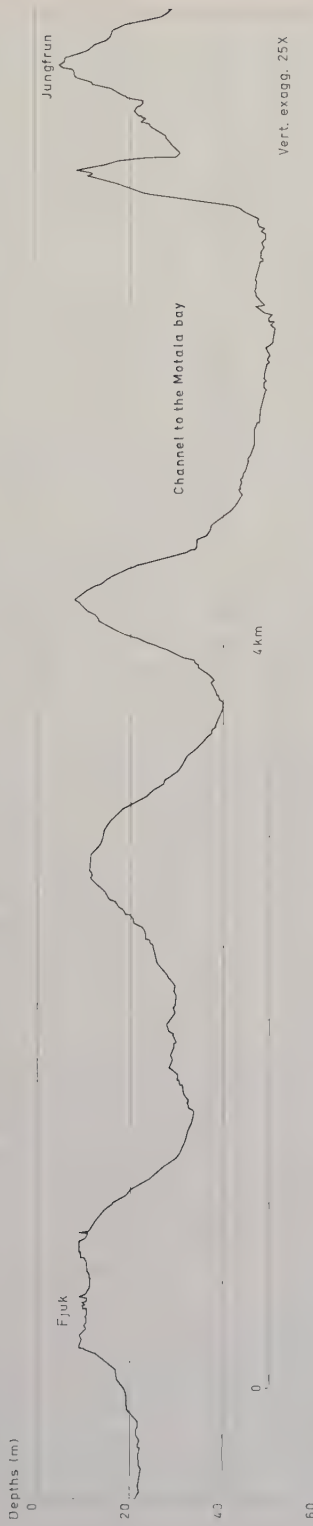


Fig. 6. Echo-sounded bottom profile along the island chain at the entrance to Motala bay. The profile was sounded close in west of the shores of Fjuk island and the Jungfrun skerry.

a granitic conglomerate by Linnarsson (op. cit., p. 26), who considered it a product of weathering *in situ*. Jönsson (op cit., p. 16) cites Linnarsson, but characterizes the rock as a coarse conglomerate or breccia.

The rock consists of large blocks, some huge slabs having a visible length of more than 6–8 m. The surfaces are rough and the shape angular. The matrix is made up of angular gravel. Especially under the water level, this gravel can be seen to form beds underneath huge blocks (examined by diving operations). The blocks have not been found to be arranged in any specific order, and even if some beds seemed to be rather horizontal, this may not be the general rule.

This material is certainly neither influenced by fluvial or glacial transportation, nor weathered *in situ*. There are no traces of the typical granitic weathering along joints, giving a regular pattern. The large range of sizes of fresh rock makes this interpretation improbable. The disorder of the blocks, their shape and the matrix argue in favour of a consolidated talus material from rockfalls along a cliff or in an open crevice. The material has probably been weathered to some extent after the deposition, and gravel has been washed down, filling the hollows. Thus, this rock should be classified as a talus-breccia. The age of its formation and its relation to the Visingsö Series is unknown.

The sedimentary rocks cover small areas around the northern section. What part they may play in the deeper parts is, however, uncertain. Some platforms are possibly covered with sedimentary rocks, but no specific subaquatic morphology can be attributed to the Visingsö Series as in the area south of this section.

The Motala bay

This section belongs entirely to the Ostrogothian Plain. The bedrock consists of Cambrian-Ordovician-Silurian sediments, resting directly on Archaean rock. The following types of sediments are represented in this series (counting from the bottom): sandstone (21–25 m), shales (28–35 m), alum shales (about 30 m), and finally limestones. The figures given in parenthesis show the thickness of undenudated parts of the series, according to Thorslund (1951). The sediments have been preserved below faults bordering the area to the north, north-west and west. The northern limit stands out in the hillside of granitic rock, running E.–W. along the northern shore of Lake Boren (north-east of Motala, cf. Pl. 1) and north of Motala. In the lowland north-west of Motala, the series borders upon coarse granite, the surface of which dips 20° towards the SE. (Linnarsson 1880, p. 15). Nearby exposed limestone beds strike N. 65°–85° W., and the northerly dip is 60°–90° (Jönsson 1887, p. 18). To the west the distribution is bounded by the ridge at the entrance to the bay, and along the shore, the Plain comes to an abrupt end at the Omberg horst.

The inclined granitic subsurface is a part of the former Precambrian peneplain. At the southern margin of the present distribution of the sediments, around Lake Tåkern, the surface lies at about 90 m above sea-level, and where lowest, in the area around Motala, at about 90 m below sea-level (Thorslund, op. cit.). The latter level is about 50 m lower than the deepest parts of the present Lake Vättern, but still higher than the low sandstone surface in Jönköping.

The nearly horizontal denudation surface of the Plain cuts the inclined beds, and thus the oldest sediments are found in strips along the southern margin, and the limestone makes up the central and northern parts. Thus, the original thickness of the limestone cannot be estimated. The largest thickness observed, around Motala, is about 90 m (Thorslund, op. cit.).

The Plain continues into the present bay without any break at the shoreline. There are only a few morphologically significant outcrops along the shore. As can be seen from Pl. 1, the bottom of the bay is very even. However, some small, flat-headed knobs, a few metres high, and with steep walls, have been found by echo-soundings. The origin of the knobs is uncertain, but their steep walls seem to preclude the possibility of glacial deposits.

A morphological feature which can be observed from the contours of Pl.1 is the rather shallow channel that from north of the Jungfrun skerry runs towards Motala. Hypothetically, this depression can be explained as being of structural origin or glacially excavated. However, it is of interest that the threshold of the outlet at Motala is not cut in solid rock. The bed of the river Motalaström consists of residual boulders on moraine and glacifluvial deposits. The depth to the limestone in the river course is not exactly known, but it is considerably deeper than in the surroundings, and from reports on work carried out on the foundations of the Motala hydro-electric power station (Kungl. Vattenfallsstyrelsen 1928) and from the list of drillings kept at the Geological Survey of Sweden, the bedrock can be estimated to lie at least 20 m below the present lake level. No statement about Pre-Glacial or Inter-Glacial levels can be based on this figure, but the shore level of Lake Vättern may have been lower than at present, and in that case it is possible that the channel of the bay formed a part of the outlet.

Quaternary geology

Introduction

The general distribution of the Quaternary deposits around Lake Vättern is shown in the maps edited by the Geological Survey of Sweden in the Aa Series. However, as has previously been mentioned, most of this mapping took place during the 19th century and the first few years of this century, and at that time little was known about the ice recession and the Late-Quaternary development of this area. Accordingly, the interpretation of the sediments is sometimes erroneous.

In agreement with J. Lundqvist (1960) the Swedish term *morän*, which includes all material assumed to have been transported and deposited by the ice, has been translated by *moraine*. Thus moraine is synonymous with the general sense of drift, till and boulder clay. Moraine forms are distinguished by compounds such as terminal moraine etc. The following lines on the modern Swedish classification of moraine have been quoted from J. Lundqvist (op. cit., p. 9).

"The moraine is the most abundant Quaternary deposit in Sweden . . . On modern Swedish geological maps it is—according to G. Lundqvist—generally classified with regard to its boulder materials as well as its fine fractions. From the boulder content four classes are obtained: moraine with low boulder content, with medium boulder content, rich in boulders and with large boulders. The fine material gives five main classes: gravelly, sandy, fine sandy, silty and clayey moraine. Combinations of these five classes are commonly used, e.g. the sandy to fine sandy moraine is the most common type. A complete characterization of a moraine must include the boulder type as well as the type of fine material. Thus twenty main classes of moraine are obtained—like sandy moraine with medium boulder content etc."

In the old maps of the Vättern area the moraine is not classified in such detail, nor are there any strict criteria of the classification.

The maps show the surface deposits generally differentiated into five genetic categories: morainic deposits, glacifluvial sediments, ice-lake sediments, marine sediments and peat. If a deposit, for example moraine, has been re-worked by wave-wash, and the resulting deposit is more than about 0.5 m thick, the sediment is classified as an ice-lake sediment or a marine sediment, depending on its height in relation to the presumed lacustrine and marine levels. A primary glacifluvial deposit, not having the specific morphology of an esker or a delta, is also generally classified as an ice-lake or marine sediment. If there is only a thin layer of re-worked material, this is characterized as a wave-washed moraine or a wave-washed glacifluvial sediment. As regards the grain-size of the sediments, three types are distinguished: gravel, sand and clay. The moraine is either classed as moraine in general (not specified) or clayey moraine.

A deposit, classed as an ice-lake or marine sediment, often inadequately expresses the sources of the present shore processes, because the surface material can be rather thin and may differ totally from the embedded sediment. The descriptions in the maps give some very valuable information about the stratigraphy observed in cuts and cliffs, but most of the old maps have very brief descriptions.

Our present knowledge of the Late-Quaternary development of the Vättern basin is derived primarily from the extensive studies of raised shorelines by E. Nilsson (1939, 1953, 1958). He has also dated the ice recession in central Sweden from varve chronology. Most of these investigations have not been published, but the results have been used by G. Lundqvist (1961) in his comprehensive map of the deglaciation of Sweden. Among older, detailed investigations in the basin, those of Munthe (1907) and Westergård (1926) should be mentioned. Palynological investigations of great importance from southern Lake Vättern have been published by Granlund (1934 *a*, 1934 *b*) and Sandegren (1951). The ice-marginal depositions in the northern parts of the basin and their relations to the changing water level have been studied in great detail by K. E. Bergsten (1943).

The following brief outline of the Late-Quaternary development of the basin is primarily based on the works of E. Nilsson and Granlund and the conclusions of G. Lundqvist. It is divided into two parts; (*a*) ice recession, and (*b*) land uplift and water levels. These introductory parts are intended to form a background to the description of the distribution of the Quaternary deposits in this chapter, the condition of the present shores, and the Late-Glacial bottom sediments of Lake Vättern, which are dealt with in a special chapter. No Pre-Glacial deposits are found in the area. Although the material has been partly re-worked in Post-Glacial time, much of the geomorphology reflects the original deposits. The description of the soils is divided into the same three geographical sections as in the chapter on the rocks.

The ice recession

The azimuthal distribution of the glacial striae (G. Lundqvist 1961) indicates a terminal ice movement in a southerly direction. Some striae of the steep borders, perpendicular to the main axis of the basin, can be judged to reflect local pressures. The striae, drumlins, terminal moraines, and other marginal accumulations all show that during the recession over southern central Sweden the ice margin had a convex shape. The topography of the landscape generally affected the configuration of the margin in detail, but where the relief had the proportions shown in the Vättern basin, the topographical effect must have been important.

When the southernmost part of the ice was still just south of Jönköping, the ice had already melted from the highest parts of the Hökensås hills up to about the latitude of the northern end of Visingsö island. An ice lake of melt water was dammed between the hills south of the basin, and there was also melt water along the lateral margins of the ice, but the highest hill tops projected above the water level (about 240–260 above the present sea-level). During the next recession, lower passes were successively exposed in the eastern and western borders, and the melt water alternately drained towards the Baltic Ice Lake and westwards, and consequently the water stage retired. When the ice margin reached the low Ostrogothian Plain about 5 km south of Hästholmen, a connection on a level with the Baltic Ice Lake was established after a lowering of the water level of the basin. According to E. Nilsson, the ice margin was still south of Jönköping about 9200 B.C., and the connection with the Baltic Ice Lake was established about 250 years later.

The position of the ice front between the border lands of the basin during this recession and the next is uncertain. A sketch map by E. Nilsson (1939, Fig. 3) shows wide areas of open water south of the margins, crossing the basin in convex narrow curves, but in another sketch map (1953, Fig. 21), almost the whole present lake area down to about Bankeryd (E in Pl. 1) is covered by an ice lobe, when the margin of the eastern border had retreated to about 10 km south of Hästholmen. In two papers, G. Lundqvist (1942 *a*, 1942 *b*) maintains that there is no glacifluvial varved clay on the bottom of Lake Vättern, but only moraine, and his conclusion is that there was a body of ice in the basin during the entire recession. However, in his map of the deglaciation of Sweden (1961), the ice front lies in a lobe off Hästholmen at the Baltic Ice Lake stage, thus leaving space for sedimentation in front of the ice.

In the period from about 8900 to 8100 B.C., the recession was slow, and there was also some advance. This is the period when the large terminal moraines (mainly glacifluvial deposits) were developed across the whole of Fennoscandia. In the Vättern area, this zone is found on the eastern border from about Hästholmen to the latitude of Stora Röknen island (P in Pl. 1), and on the western shore from the latitude of the Flisen shoal (J in Pl. 1) to the low land north of Karlsborg (cf. G. Lundqvist, 1961).

In the year 8223 B.C., the Baltic Ice Lake was lowered to the level of the sea by an outflow through the Örlen valley south-west of Karlsborg. This exact dating is derived from the varve chronology, established by E. Nilsson (personal communication, 1962). The lowering was possible because of the ice-free Örlen valley, and the ice recession from the northern slope of the dam formed by Billingen hill, about 40 km west of Karlsborg.

At the early stage of the Yoldia Sea, the ice margin was well north of the Motala bay, and probably just south of Stora Röknen island (K. E. Bergsten 1943, Fig. 78). A hundred years later, certainly no active ice was left in the basin (G. Lundqvist 1961, the map).

The land uplift and the water levels

On the assumption that there is a linear relationship between the amount of Post-Glacial uplift and the distance along lines perpendicular to the isobases, any synchronous, raised shoreline in this direction is a straight line. With a decreasing rate of uplift, the angle between shorelines at equal intervals of time decreases. On the basis of these assumptions (not expressly stated anywhere; cf. Norrman 1964), E. Nilsson (1953) has plotted the raised shore scarps of the Vättern basin on a vertical plane along the basin, and connected the scarps falling along straight lines. The positions and the heights of the shore scarps were obtained from extensive field work. From these reconstructed, raised shorelines and the position of former outlets, he has evaluated the succession of ice lakes and the marine and fresh water stages of the basin.

The oldest and highest marks of dammed melt-water surfaces have a steep slope towards the south, the gradient gradually decreasing for lower, more recent shorelines. As regards the scarps older than the Baltic Ice Lake stage of the basin, it may be questioned, whether they reflect coherent water surfaces along the lateral ice margins (Norrman 1963 *a*).

For the period from the Yoldia stage to the present time the gradients of the shorelines can be fitted to an exponential function, the best fit being obtained by the function

$$\alpha_t = 793.33\ 10^{-4}\ e^{-2.4290\ 10^{-4t}} \qquad (1)$$

where α_t is the final gradient of a shoreline developed at the time t, and $t = 0$ at the birth of Christ (Norrman 1964).

As has already been mentioned, the highest marks of dammed water around the southern end of the basin are found at levels about 240–260 m above sea-level. The first level, belonging to the Baltic Ice Lake stage, lies at about 126 m above sea-level at Jönköping (37 m above the present lake level), and at about 147 m at Hästholmen. Immediately before the lowering to the sea-level, the levels at the same points were about 100 and 131 m above sea-level (cf. Fig. 7). The outflow to the sea through the Örlen valley in 8223 B.C. lowered the water 26 m (Westergård 1926, p. 67, G. Lundqvist 1931, p. 127). The water level, equal to that of the Yoldia Sea, is found at Karlsborg at the height of 126 m, at Gränna at 92 m, and, extrapolated to Jönköping, at 78 m above sea-level (11 m below the present lake level).

During the marine phase the change of water level was dependent on the land uplift and eustatic changes of the sea. The uplift was rapid, the passes of the Karlsborg area and the Ostrogothian Plain were lifted above the sea, and the only connection left was through the northern archipelago and past Askersund to a threshold at Lake Tibon. The basin was isolated from the equivalent level of the sea at this outlet about 7700 B.C. (according to equation 1). The Baltic was still connected to the sea through the broad lowland north of the bordering hills of the Vättern basin. At the time of the isolation, the water level in southern Vättern reached its minimum, the height at Jönköping being estimated at about 46 m above sea-level (43 m below the present lake level), at Motala at 84 m, and at Askersund at 102 m above sea-level. Trough the isolation, the lake level became independent of the sea, and the land uplift, being more rapid towards the north, caused a reverse, submerging motion of the lake level. When the lowest pass of the Ostrogothian Plain was flooded, a new outlet was cut about 5700 B.C.,and this outlet is the one still existing at Motala. Because of the change in position of the outlet, the unequal uplift since then has caused a submergence south of Motala and an emergence north of Motala.

The rate of submergence around the southern end of the lake from about 2300 B.C. to the present time has been determined by pollen analysis of submerged peat layers (Granlund 1934 *a*, Sandegren 1951). Some additional help has been given by carbon-14 datings and calculations of the present rate of submergence

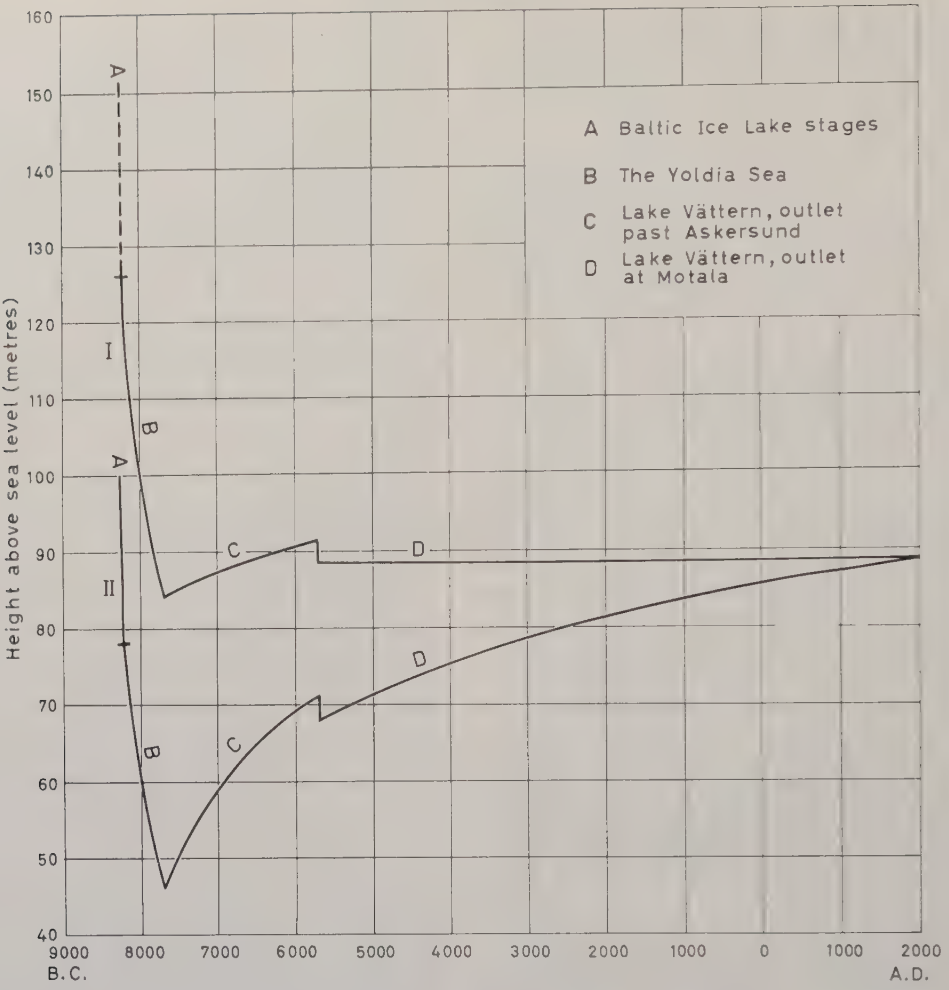


Fig. 7. General trend of the water-level changes in the Vättern basin from 8300 B.C. to the present time; I, at the latitude of Motala; II, at Jönköping (mainly according to E. Nilsson).

(Norrman 1964). The present rate of submergence in Jönköping has been found to be about 11 cm per 100 years.

The deposits of the central and southern section

With respect to height, the distribution of morainic and glacifluvial deposits is rather unusual in this section. The upper parts of the Hökensås hills are occupied by large fields of glacifluvial sediments, constituting a complex pattern of eskers, kames and deltas, while, on the other hand, much of the lowlands is covered with clayey moraine.

The sediments of the upper parts of the Hökensås hills originate from the time when the whole basin was filled with glacier ice up to the margins of the high borders. The subglacial drainage system brought the material to the exposed areas of Hökensås. As can be judged from the distribution of the material (the main feeding eskers running up the westerly hillside about 10 km south-west of Hjo), much of the melt water did not emanate from the Vättern basin. The glacifluvial material is found up to the highest levels of the dammed waters. The highest parts, which projected above these levels, are covered with moraine.

In the highland east of the basin, coarse glacifluvial material is not so abundant, most of the hillsides and small valleys having moraine deposits. However, an esker runs in the valley north of Lake Landsjön, and varved clay also settled in this area. A drilling at the parish church of Skärstad, 2 km east of Lake Landsjön, and at a height of about 195 m above sea-level, showed 22 m of clay on 4.5 m of gravel (Munthe and Sandegren 1951, p. 59).

During the lowering of the ice surface, the melt water re-worked the glacial and glacifluvial deposits along the ice margins. In the steep cliffs of the eastern border very little material was left.

Thick deposits again appear along the western shore at the levels of the Baltic Ice Lake stages from about the latitude of the northern end of Visingsö and southwards and also at the southern end of the lake. Along the western shore, they constitute the platform bordering the hillside.

In the description in the geological map it is said that the sandy ice-lake sediments form flat or slightly undulating fields, the sand being from fractions of a metre to several metres thick, as exposed along the shore, and in some deeply cut ravines more than 10 m (Munthe and Sandegren 1951, p. 60). This sand embeds varved clay and boulder clay. These figures are actually minimum figures for large areas. At the mouth of the broad valley at Bankeryd (E in Pl. 1), the thickness of the sandy sediments is generally more than 16 m, the shore cliff having a height of 14–18 m.

In the map description cited, folded structures, interpreted as being the result of ice pressure, are described (op. cit., p. 58), and in the area south of Bankeryd (near Jönköping), E. Nilsson (1939, Fig. 4) has observed morainic material deposited on ice-lake sediments. Several more indications of ice advance during the recession in the central and southern basin will be mentioned later. However, the general term "ice-lake sediments" gives a very vague idea of the origin of these deposits. By examination of the shore cliffs, it has been found that the general sequence from bottom to top is as follows: clayey moraine, glacial clay (not always varved), silty sand, and sand with layers of gravel or even pebbles.

The silty sand layers often have a folded or convolute structure (Potter and Pettijohn 1963, p. 152), and these folds are not generally of tectonic origin (from ice pressure), but reflect a turbid transportation or slumping (Figs. 8 and 9). Most of the silty sands, like the pure sand, are not, as at somewhat higher levels, the product of re-worked glacifluvial material from any type of shore processes, but primary fluvial sediments, which have either been subglacially deposited or deposited between the hillside and the margin of a low glacier at the time of the Baltic Ice Lake stages.

As regards the present shore processes, the small amount of coarse material and the position of the upper surfaces of the cohesive deposits, from a few metres above to below the lake level, are of great importance.

The promontory of Svedudden (E in Pl. 1, see also Figs. 10 and 74) seems to have had a special origin. Most of the material lies in the sand to medium gravel range, with a high content of quartz. As can be judged from the sedimentary structure exposed in the large central sand pit and the surface morphology, this deposit originally had the shape of a deltaic fan, facing the basin. It may partly have been formed subglacially. During the falling water levels, the rivulet draining this part of Hökensås cut changing stream channels in the sediments. Two channels, shown in Fig. 10, have been marked by hatched, open arrows.

According to E. Nilsson (1958, Fig. 8), the lowest well-marked shores of the Baltic Ice Lake stages in this area lie at about 120 and 114 m, levels which can be connected with the abrasion surface found in the same interval in Fig. 10.

At the southern end of the lake between Jönköping and Huskvarna a high promontory, called Rosenlunds bankar, stands out (B in Pl. 1, see also Fig. 74). This is a mainly glacifluvial, deltaic deposit, severely abraded in recent time. It has the form of a plateau along the shore, the flat top having a height of 124–129 m above sea-level (35–40 m above the present lake level). The inland slopes are gentle, the dip ranging from about 2° to 4°. The shore is a steep cliff (Fig. 11). Offshore, a shallow triangular platform is found.

The position of the formation between the valleys of Jönköping and Huskvarna, its level, and the structure of the sediments, all indicate that the deposition took place during a temporary advance of the ice.

The following main sequence of sediments is exposed in the central parts of the cliff and

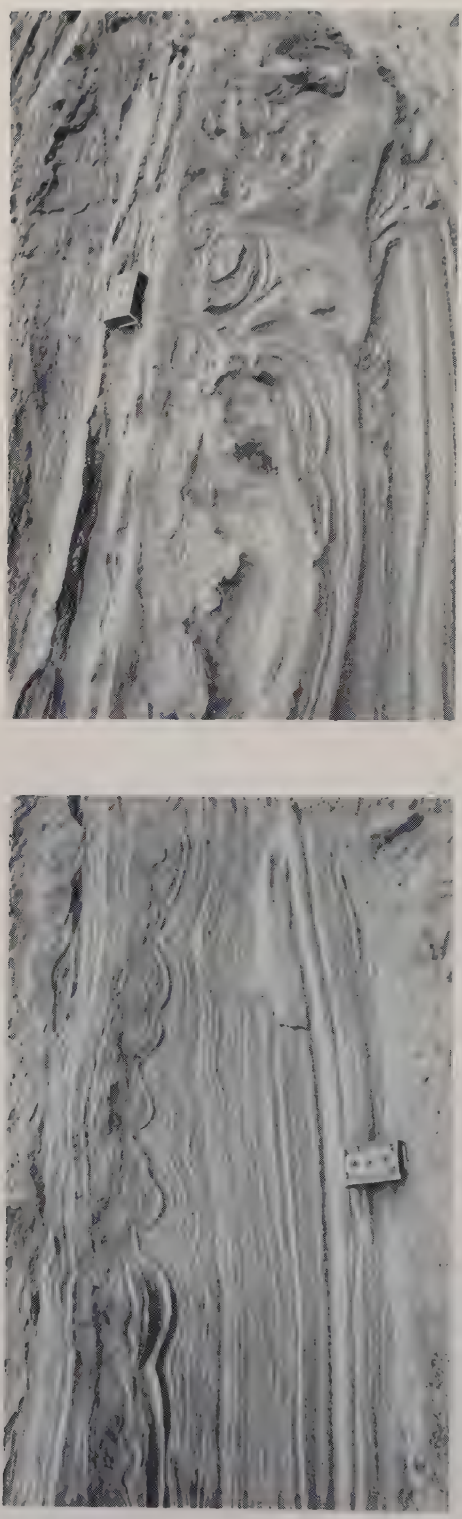


Fig. 8. Convolute lamination in glacifluvial sand beds exposed in the cliff north of the outlet of the river Dumneån. The left-hand photograph shows embryonic "convolutes". The right-hand photograph, taken 4 m north of the former, shows a well-developed structure. Ordinary match box used as a scale, Oct. 1962.

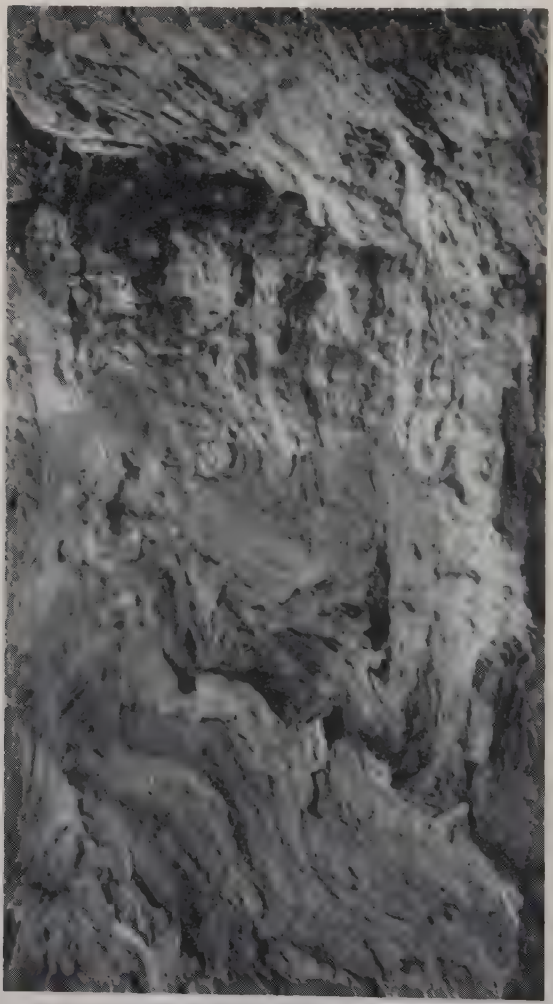
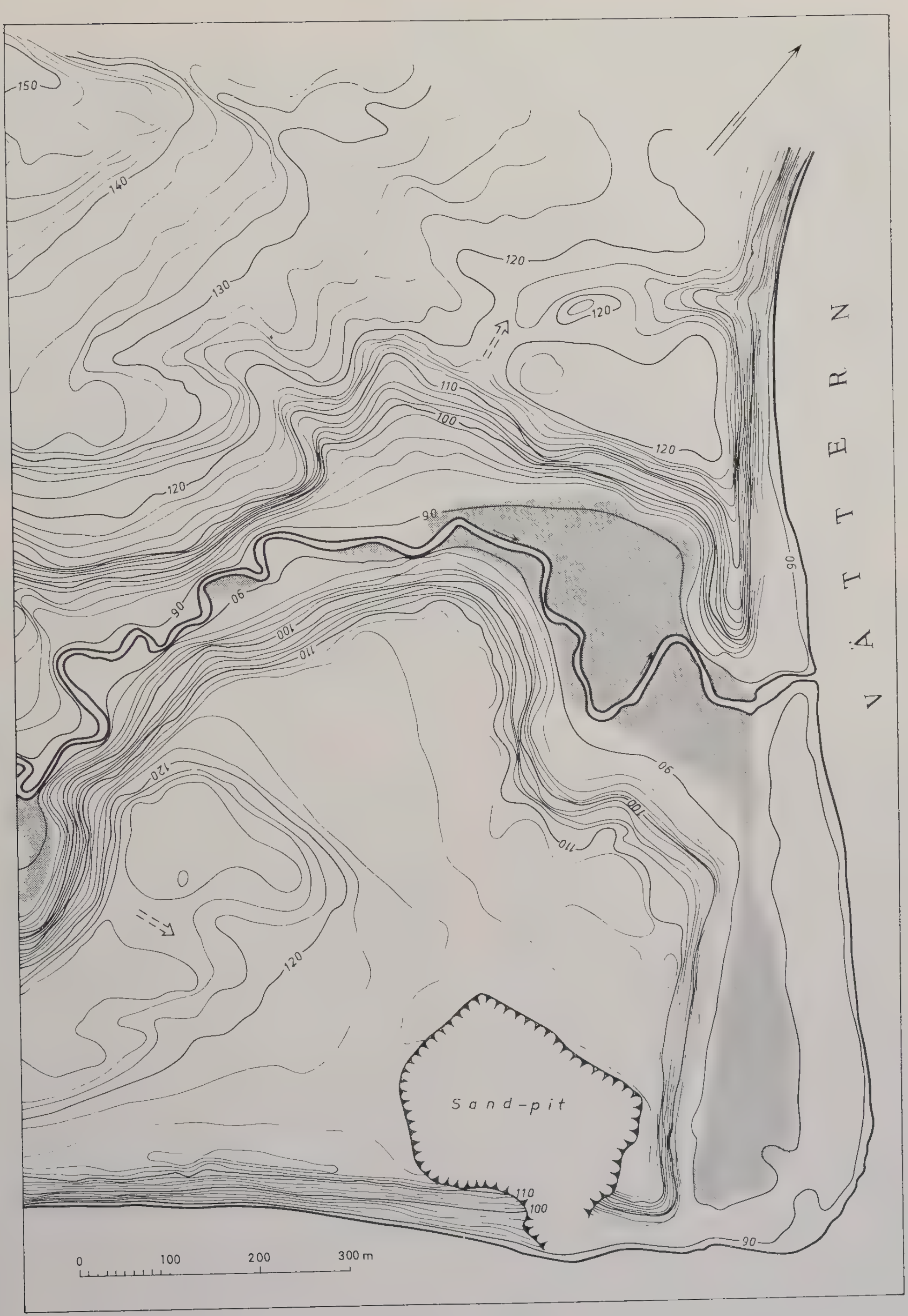


Fig. 9. Slump structure of silty and sandy glacifluvial beds exposed in the cliff of the Vilhelmsro area. Height of photograph about 0.8 m. Aug. 1961.


150
140
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90
90
90
100
110
90
90
120
100
110
120
Sand-pit
110
100
90
VÄTTERN
0
100
200
300 m

Fig. 10. Photogrammetric map of the outer part of the Svedudden promontory. Contour interval 2 m. Heights above sea-level. Bogs shaded. Old channels marked by open arrows. On top of the south-eastern cliff a dune ridge is enclosed by the 116 m contour.

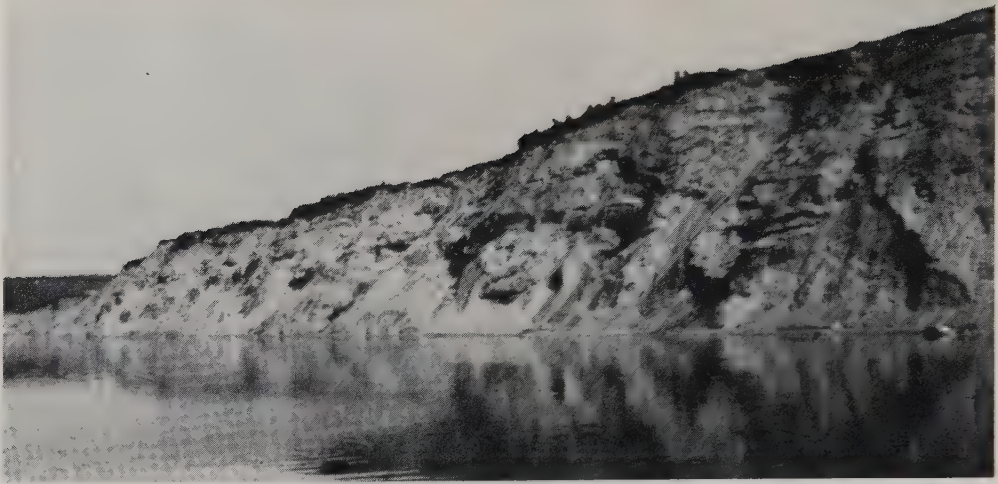


Fig. 11. The eastern part of the Rosenlund cliffs. Height of the nearest part 36 m. Talus cones of sand cover the cohesive sediment of the lower cliff. July 1962.

on the subaquatic platform; glacial grey clay, well-sorted sand rich in quartz, sand and gravel, partly rich in material of the Visingsö Series, and finally eolic sand.

The clay is silty (cf. analysis no. 108/61), and in some parts scattered pebbles are found. It is very homogeneous, and there are no traces of varves. Munthe (1907, p. 134–135) regards most of this clay as being "ice-lake marl", but the term "morainic marl" is also applied. The grain-size distribution indicates water transport of all the clay. This transport was most certainly subglacial, as the deposition may also have been. The same clay is found on the shallow bottom offshore. The upper boundary is an erosion surface. Water flow has eroded deep, V-shaped stream channels, which, on examination, seem to be mainly directed towards the SSE. A proof of this erosion has already been given by Munthe (op cit., p. 139), who mentions small balls of the clay found in the covering sand beds. Through this erosion, the height of the clay surface varies from about 3 m to 15 m above the lake level. The clay is tectonically structured by compression in conformity with its surface.

The well-sorted sand filling the channels has well-developed ripples and foreset beds. The upper limit of these beds is rather horizontal, as is seen along the cliff at a height of about 21 m above the lake level. In the covering beds of coarser sand and gravel, topset bedding predominates. The sorting is good within the single layers, but the changes in grain size are very frequent and rapid.

The eolic deposit at the top is of recent origin, the sand being blown from the naked cliff. The material partly forms a small dune ridge, up to 3 m in height, at the edge of the cliff. I have observed how, during only a few weeks, 1–5 cm of sand can be deposited on the fields behind the cliff.

The material on the eastern and western flanks of the plateau is of a different character. The eastern flank consists of a clayey moraine with a comparatively high content of boulders and large blocks (Fig. 12). The embedded sediments are very disturbed at the contact with this moraine. The content of coarse material is highest in the easternmost part. From the eastern end of Rosenlunds bankar a low ridge of boulders and blocks, called Stenrevet (the stone reef), runs on the bottom all along the north-eastern margin of the subaquatic platform. The whole platform is strewn with boulders and blocks. By examination in diving operations, it has been found that this material, mainly consisting of Archaean rock, is generally rather well-rounded, but there are also pieces with a distinctly angular shape. The layer seems, as a rule, not to be very thick. The subsurface, always clay, is often exposed between the boulders, and here and there are areas of some tens of square metres without

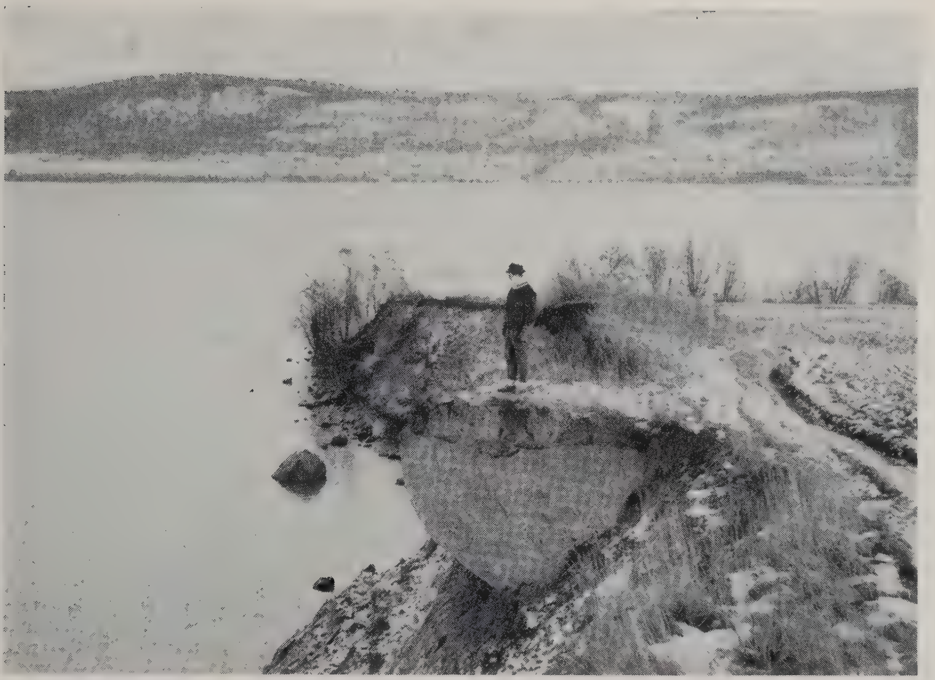


Fig. 12. The eastern end of the Rosenlund cliffs. The large block marks the western limit of moraine pushed up on the glacifluvial sediments. April 1959.

any covering material. In a zone near the present shore, the boulders are very few and scattered. This zone has a width of 60–150 m with a general broadening towards the west. Nowhere has a boulder been found embedded in the clay. These boulders must be regarded as residual material, originally belonging to levels far above the present bottom. They can be considered to have been left as an erosion pavement during a continuous abrasion still in action.

The western flank also consists of morainic material, but in this part few boulders are found. Munthe (op. cit., p. 133) classed most of this material as morainic marl and ice-lake marl. In my opinion it is better described as a moraine, rich in silt and sand, with enough fine material to give it a cohesive character. The water content must have been high enough to make it very mobile when transported, and it was probably in a state of flow until deposited. With a thickness of up to 3 m, this type of material also covers areas of the central part of the plateau.

In the fine sediments at the base of the sand layers, fragments of plants have been found, but only a leaf of *Salix polaris* has been identified (Munthe 1907, p. 130). Allvin (1839) mentions a find of mollusc shells in the cliff. However, this observation is uncertain, and no shells have been found since.

From present knowledge, the development of this deposit may be interpreted generally, but cannot be stated with certainty. The contact between the lowest exposed, fine sediment and the sand indicates a discordance, due to an ice recession, followed by a temporary advance. The amount of sediment eroded before the deposition of the sand cannot be determined. The eroded clay surface must have been rather hard, because there is no zone of transition between the fine sediment and the sand.

The morainic cover of the flanks and the topset beds of rather coarse material make it probable that the erosion and the deposition took place during the advance of a low ice tongue, which was pushed into the valleys of Jönköping and Huskvarna, but, because of the bottom topography, left a gap in the area of the Rosenlunds bankar. From the rate of the present abrasion (Fig. 124) and the morphology of the subaquatic platform, it seems most probable that the original formation covered this platform entirely. The granitic boulders and blocks of the platform, being most frequent along the north-eastern margin, make it probable that material previously deposited in the southern part of the basin was pushed up at the ice front. Much of the glacifluvially transported sediments, exposed in the cliff, may have originated from such accumulations. The flank depositions reflect the conditions in areas of less intensive water flow.

The period best fitting this interpretation is the early Baltic Ice Lake stage of the basin, when the level of the marginal melt water was lowered to an estimated height of about 126 m. This lowering caused a rise in ice pressure at the margins, and during the following period (8900–8500 B.C.), when there was a general halt in ice recession, advances within the basin are most probable.

At the part of the Jönköping bay west of the outlet from Lake Munksjön, sandy shore sediments from the Baltic Ice Lake stage form layers with a thickness of several metres. The area east of this outlet is a bay mouth barrier, continuously formed during the present period of submergence. Through this barrier, the small lakes of Munksjön and Rocksjön have been enclosed in a formerly open bay of Lake Vättern. The valley of the river Huskvarnaån has also gradually been filled up with sediments during the submergence.

In the main part of the southern and central section, not yet described, clayey moraine predominates. On the narrow platform north of Huskvarna the content of boulders is rather large in the deeper parts of the moraine. The surface is covered with sandy sediments accumulated by wave-wash.

In the lowland at Gränna the frequency of boulders is generally low. The moraine of the lowest part near the shore is covered with fine sediments, "ice-lake clay", of variable thickness, generally not thicker than 0.5–1.0 m. Its sedimentary structure has been destroyed by plant roots and weathering. The total thickness of the deposit also varies. In the northern part, where there are shale cliffs, the thickness is less than 3 m, but in the area near the Gränna harbour it can be estimated to be about 5–6 m.

On Visingsö, clayey moraine is exposed in the cliffs all around the island. This moraine has a high content of the local rocks. Where the bedrock is exposed at the base of the cliff, a bottom layer of broken-up pieces of the bedrock can be observed. This layer has a thickness of 0.3–1.0 m. Cobbles of the Visingsö Series and pegmatitic rock, found at somewhat higher levels, are generally extensively weathered. The scattered large boulders and blocks are almost exclusively granitic.

The relation between the rocks of the Visingsö Series and Archaean rock in the grain-size intervals 100–20 mm and 20–10 mm (converted from square mesh) is shown in Table 1. The table is based on 10 samples, each weighing 10 kg, taken at a height of 1.0–1.5 m above the base of the shore cliff on the northern shore. In this area, the bedrock is shale. 94 % of the material in the whole interval 100–10 mm was of the Visingsö Series.

Table 1. Percentual grain-size distribution of Visingsö moraine with regard to crystalline and sedimentary rocks.

Particle size (mm)	Crystalline	Sedimentary
> 20	0.98	14.89
10—20	0.23	3.80
5—10	0.23	3.97
< 5	75.90	

As regards grain-size, 25 samples from the western shore have shown that less than about 0.5 % is coarser than 100 mm, and about 45 % is coarser than 0.2 mm, of which medium sand (0.2–0.6) forms about 23 %.

The total thickness of the deposits near the shores is generally about 3–5 m. From well-drillings, the maximum thickness can be estimated to be about 8 m.

Along the crest of the central part of the island, at a height of 25–33 m above the lake level, the ground is covered with sand and gravel. This gravel is, according to Munthe and Sandegren (1951, p. 60), a product of a wave-wash, which has affected the "ordinary clayey moraine of the area". I have examined a few shafts near the crest. In these shafts, which had a depth of 2–2.5 m, the gravel was lying in horizontal, rather indistinct beds all down to the bottom. The space between these larger particles was filled with poorly sorted sand and silt. The origin is uncertain, because a thickness of this size requires very large volumes of "the ordinary clayey moraine", volumes which

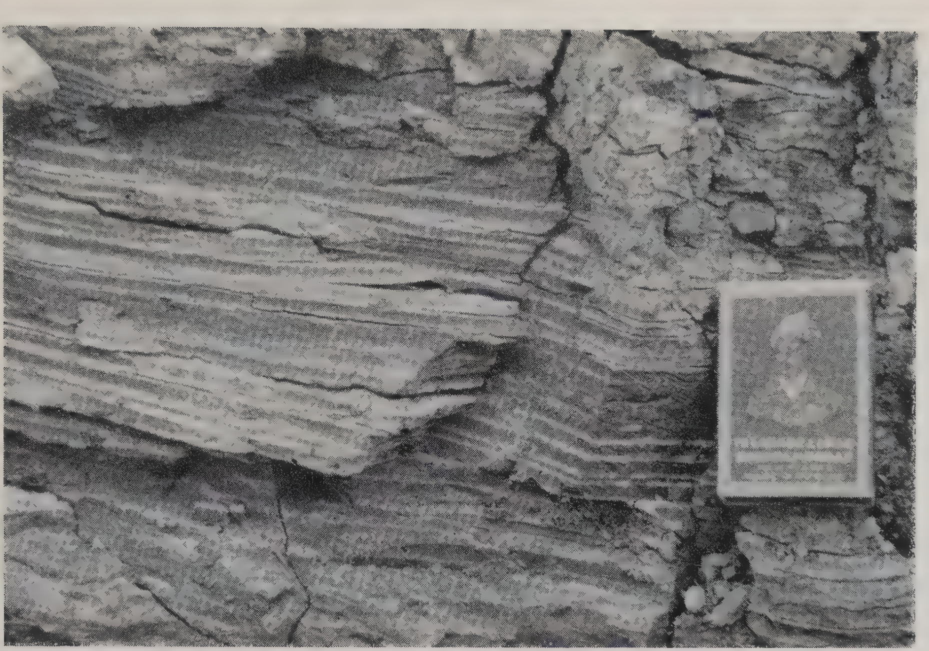


Fig. 13. Top bed of varved clay in cliff on the western shore of Visingsö island. Ordinary match box (5.3 × 3.7 cm).

can hardly have been available in the highest part of the island.

The moraine of the lower areas, that is, the northern part of the island and along the western shore, is, as in the Gränna area, covered with a brownish, sedimentary, silty clay. As can be observed at the shore cliff, this cover is often very thin, only having a thickness of one or a few decimetres. The moraine surface is not totally flat, and in cavities the clay is thicker. The maximum thickness is about 1.5 m. Where the clay is thick, the sedimentary structure is preserved, and the character of a varved clay is obvious(Fig. 13).

Besides this covering of varved clay, remnants of an older sedimentation are found, embedded in the moraine (Munthe and Sandegren 1951, p. 57). This sediment is also a varved clay, which I have found to exist at localities all along the northern and western shores of the island. The largest, exposed, coherent parts have a maximum length of about 30 m. The thickness is very seldom more than 1 m, usually less than 0.5 m. The layers are partly deformed by mangling and tearing, but are generally not much folded. The lowest parts of the sediment regularly lie about 0.5–1.0 m above the bedrock surface, that is to say, immediately above the actual local moraine with coarse material.

This clay has a specific stratigraphy and colour. The varve structure is made up of chocolate-brown strata with a maximum thickness of about 1 cm, and intervening, light-greyish-brown, thin layers (0.5–1.5 mm). The boundaries between the strata are very distinct. The chocolate-brown clay is very fine-grained with about 95 % $< 2\ \mu$ (cf. analyses nos. 71 and 73/61). The thin strata hold coarser material, about 45 % being $< 2\ \mu$ (cf. analysis 72/61). The sediment does not react to hydrochloric acid.

The presence of this embedded sedimentary clay and the varved clay on top of the moraine demonstrates that the basin was once free of ice with bottom contact to at least north of Visingsö, then the ice advanced to at least south of the island, and at the final recession, the fine glacial sediments could again settle in the southern part of the basin. The lack of a cover of fine sediments in higher but very distinct areas in the southern part of the basin, as, for example, in the valley at Bankeryd, indicates that the last sedimentation took place during a stage of rather low water level. Munthe and Sandegren (1951, p. 62), on the contrary, suggest that the upper clay at Visingsö was deposited before the Baltic Ice Lake stage of the basin, an interpretation which seems most improbable.

Along the western shore of the mainland north of Visingsö, no glacifluvial sediments are found until the plain south of Karlsborg (the area below the 100 m contour in Pl. 1). Moraine predominates in the lowland (Guldkroken) south of that plain. The moraine is to a large extent covered with re-deposited, sandy sediments from the lowest stage of the Baltic Ice Lake and from the shores of the Yoldia stage. Near the surface, the moraine of the area south of Hjo is almost free of boulders and blocks, but, as can be seen in rill cuts and at shore cliffs, the content of coarse material is often high at lower levels. North of Hjo, the coarse part is larger, and residual boulders are quite frequent in the wave-washed areas.

The deposits on the hillside bordering Guldkroken have a morphology resembling a hummocky moraine terrain, as defined by Hoppe (1952, p. 54). Above the lowest level of the Baltic Ice Lake stage, there are almost no traces of a morphology, which can be connected with large open waters (cf. Norrman 1963 *a*).

In contrast to Guldkroken, most of the superficial deposits on the plain south of Karlsborg are of glacifluvial origin. There are stumps of eskers and wide fields of glacifluvial gravel, sand and silt. Varved clay is found on the lower parts south-west of Karlsborg (cf. Westergård 1926). At Karlsborg, "Late-Glacial clay" has also been found under "mighty beds of sand" (Westergård, op. cit., p. 60).

The Motala bay

The Ostrogothian Plain lies within the stagnation zone of the ice recession. There is a large system of terminal deposits south of Lake Tåkern, and another runs from Vadstena to the northern foot of Mount Omberg (cf. Lundqvist 1961). Evidence of ice oscillations is found in the low hills at the northern end of the peninsula north-west of Vadstena and at the Sandön islet (N in Pl. 1), where glacifluvial beds of sand are embedded in morainic deposits (Jönsson 1887, p. 27). Large fields of sandy, glacifluvial deposits are found in a course along the shore north of Motala and past the town towards the SSE.

Between these morphologically well-defined deposits, the ground is covered with a mosaic pattern of clayey moraine, morainic gravel, sand, silt and clay (cf. the geological maps, SGU Series Aa, Nos. 102 and 130). Near the shore, the thickness of these deposits seems to range from about one to some few metres. Blomberg (1905, pp. 35–36) gives the following descriptive account of the distribution of the varved clay at the southern part of the plain. "The varved clay appears in many places, however, generally only as small plots at the surface; nevertheless its distribution within this area is not insignificant, as the clay can be followed under the sand of the Skeninge plain and at the bottom of the Dags moss. Probably, the clay also covers parts of the bottom of Lake Tåkern. The clay is found chiefly in depressions of the clayey moraine and is less frequent in areas of morainic gravel, which generally appears at higher levels. The large deposits of varved clay seem to fade out at a height of 110–115 m above sea level." The same pattern is found in the northern part of the plain (Jönsson, 1887), where the varved clay is generally hidden by "more recent sediments".

The deposits of the small islets and skerries in the northern part of the Motala bay are sparse, mainly consisting of residual gravel and boulders left by the swash.

The northern section

This area has, as regards glacifluvial sediments, been characterized as a region of valley fillings near the highest shore (K. E. Bergsten 1943, p. 185). The deposits of this section belong to the end of the Baltic Ice Lake stage and the early Yoldia stage. After the stagnation in the plains, the ice recession in this area was very rapid. It meant fast melting and large glacifluvial flows. "The glacifluvial deposits were formed in the marginal zone of the oscillating land-ice. They are here situated in the valleys as oses and subaqueous valley trains near the highest raised wave-cut beach. They are also formed as large lateral terraces, and as subaqueous marginal deltas. In high-situated areas some metres above the highest raised water-level they may form small supra-aqueous out-wash plains. The topography of the ground has greatly influenced the morphology of the fields" (K. E. Bergsten 1943, p.

243). The most impressive deltaic deposits are found just north of the Motala bay (Djurkälla fältet) and at the margin of the plain north of Karlsborg (Perstorps fältet).

These glacifluvial deposits are without doubt the characteristic feature of the section, but a very small part is situated in direct contact with the present shore. Only about 5 km of the shores are cut in glacifluvial deposits. There is one area at the plain opposite (north of) the Karlsborg peninsula, and one esker runs along the north-western shore of the island of Lilla Röknen (north-west of Stora Röknen) and along the mainland north of that island. Finally, glacifluvial sediments are found on the eastern shore of the Hammar peninsula. From the geological map (Blomberg 1901), the central and southern parts of Stora Röknen island could be expected to have been originally covered with glacifluvial deposits, because this area is mapped as "marine sand". However, this mapping is erroneous. The central part consists of morainic boulder and block ridges. A large part of the southern peninsula is covered with shingle beach ridges (stentorg), and only its south-western part is covered with sand, mainly eolic deposits.

Most of the shores of this section are rocky, and the areas between the rock outcrops are covered with sandy moraine or morainic gravel and beach sand. Some varved clay is found in the valleys and the plains north of the archipelago. The upper height limit of the clay lies at about 120 m (Blomberg 1901, p. 33).

Climate

Introduction

Southern central Sweden belongs to the vast region with a humid continental climate in the northern mid-latitudes, which has cool summers and a snow cover in winter. Its vegetation is characterized by northern spruce and fir.

Apart from the winds, the climatic component of greatest importance to the shore is the temperature because of the processes associated with frost and ice.

Besides the published statistics, the author has made use of diaries and wind records from meteorological stations kept at the Meteorological and Hydrological Office in Stockholm. The wind statistics have been difficult to handle because of the situation of the meteorological stations and the character of the observations.

Temperature

The summer, defined as the period with a daily temperature above 10° C, lasts from the 20th of May to the 20th of September. The winter, with a daily temperature below freezing-point (0° C), lasts from the 20th of November to the 20th of March. (All figures of temperature refer to means of the period 1901–30, according to Ångström 1938, 1953, unless otherwise stated.) The number of days with frost, defined as days when the minimum temperature has fallen below −0.1° C, is on an average 130 (Atlas över Sverige 1953, Sheet 27–28), and extreme winter, with a mean temperature below 10° C, only lasts for 10 days.

The temperature within the area varies according to the height, latitude and proximity to Lake Vättern. The pattern can be seen in the monthly distribution, exemplified by three meteorological stations (Table 2): Askersund, situated at the northern end of the lake at a height of 96 m (8 m above the lake), and sheltered from the large open-water surface by the archipelago with narrow inlets, Jönköping, at the exposed southern shore and at 97 m (9 m above the lake), and finally Flahult, situated in the boggy highland 9 km south of Jönköping at 224 m above sea-level (136 m above the lake).

The difference in winter temperature between Askersund and Jönköping can, as to two-

Table 2. Air temperature in the period 1901-30 (°C).

Station	Lat.	Height	1	2	3	4	5	6	7	8	9	10	11	12	Year
Flahult	57° 42′	224	—3.2	—3.4	—1.0	2.9	8.3	12.2	14.8	13.1	9.6	5.3	0.8	—2.0	4.8
Jönköping.....	57° 47′	97	—1.5	—1.7	0.5	3.9	9.2	13.6	16.2	14.7	11.1	6.9	2.5	—0.3	6.3
Askersund	58° 53′	96	—3.0	—3.0	—0.6	3.8	9.8	13.9	16.6	14.6	10.6	5.7	1.0	—1.9	6.1

thirds, be ascribed to the different latitudes with respect to the average variation with latitude (Ångström 1938), and as to one-third to the fact that the southern part of the lake is often ice-free, but the archipelago is frozen. The differences between Jönköping and Flahult are mainly due to the different altitudes, but in autumn and winter the effect of Lake Vättern on the temperature in Jönköping is obvious.

Precipitation

The precipitation area of the Vättern basin is 6360 km². About one-third (35.7 %) is taken up by lake surfaces (Melin 1955, p. 217). According to SMHA (1946, Table 3), the area of Lake Vättern, including the connected Lakes of Munksjön and Bottensjön, is 1912.4 km², which corresponds to 30.1 % of the total precipitation area. The bordering land is only, on an average, about 15 km. With this distribution of land and water the estimation of the total precipitation of the area must be uncertain, as the figures for the Vättern surface must be extrapolated from stations situated around the lake. The uncertainty is augmented by the differences in level of the border land (cf. Pl. 1). The average height of the area is 145 m (Melin 1955, p. 217), while the surface of Lake Vättern lies at 88.5 m.

As for the rest of Sweden, except some local maritime regions, the seasonal distribution has one maximum, occurring in late summer. This maximum can chiefly be ascribed to the convectional heaving of air masses during the warmest period.

The annual precipitation in Sweden has been calculated by C. C. Wallén (1951) on the period 1901–30 and by F. Bergsten (1954) on the period 1921–50. The two investigations thus have 10 years in common. In Table 3, the means of four stations west of the lake and four stations east of it are tabulated together with the precipitation recorded on the island of Visingsö (1938–50).

Table 3. Precipitation in Lake Vättern

Locality	Period	mm
W. side	1901—30	665
W. side	1921—50	650
Visingsö	1938—50	431
E. side	1901—30	580
E. side	1921—50	575

The differences between the two periods are small, and the figures clearly show the influence of the different roughness of the western hills and the lake surface.

Wallén and Bergsten have illustrated the areal distribution by isohyetic maps. In the Vättern basin the courses of the isohyets of the two periods have been somewhat differently drawn over the lake surface. From these maps the mean precipitation (in mm) of the total precipitation area (I), of Lake Vättern (II), and the area around the lake (III) has been calculated (Table 4). As can be seen from the table, the differences between the two periods are insignificant.

Table 4. Precipitation (mm) in the Vättern basin.

Period	I	II	III
1901—30	593	551	611
1921—50	594	554	611

15–20 % of the precipitation falls in the form of snow (Atlas över Sverige 1953, Sheet 31–32). The part which falls over Lake Vättern is to a large extent immediately transformed into water, as the main part of the lake is often ice-free (cf. Fig. 28).

The normal variations in annual precipitation in southern central Sweden are among the lowest in the country. The mean deviation calculated on the period 1901–30 was about 11 %, which corresponds to about 70 mm of precipitation (C. C. Wallén, op. cit.).

Evaporation

On a map showing the distribution of evaporation in Sweden Melin (1949, Fig. 2) has placed the figure 360 mm immediately west of Lake Vättern. F. Bergsten (1950) has stated that the value 360 mm should be increased by about 10 % in the central parts of southern Sweden. The figures given by Melin and Bergsten are based on relationships between precipitation and run-off in certain, specially investigated, river systems.

Calculations concerning the humidity (*H*) in Sweden below the mountain range on the Norwegian border have been carried out by Tamm (1959). He defines the humidity (*H*) as the amount of precipitation (P) which escapes the evaporation (E). The calculations are based on a linear association between mean annual temperature (T) and evaporation:

$$E = C_1 + C_2T \tag{2}$$

and $H = P - E$.

The constants C_1 and C_2 are determined from mean temperatures of the period 1901–30 (Ångström 1938), precipitation of the period 1921–50 (F. Bergsten 1954) and run-off values from Melin (1955) and the Meteorological and Hydrological Office. The result of the investigation is shown in a map with isolines for every 50th mm of "residual precipitation" (op. cit., Pl. 2). It is based on the areal distributions of the precipitation and the association

$$E = 220.9 + 30.4\ \mathrm{T}. \tag{3}$$

Within the Vättern basin data from 17 meteorological stations have been used.

From this map, the residual precipitation of the Vättern basin has been calculated by the present author, giving a value of 227 mm. The surface of Lake Vättern was excluded, as the association between evaporation and temperature can only be considered as valid for a "normal" distribution of land and lake surface. With a run-off equal to 227 mm and a precipitation of 611 mm (Table 4), the evaporation of the area around Lake Vättern should be 384 mm per year. If this value is inserted into Tamm's formula, we obtain a mean temperature of 5.4° C, a value which agrees well with the actual mean temperature. With regard to the character of the calculation, this evaporation value (384 mm) cannot be considered as significantly deviating from the previously mentioned 360 or 395 mm.

The balance between precipitation and evaporation on the surface of Lake Vättern has been estimated on the assumption that the mean run-off from the total precipitation area is equal to the amount of water obtained from the surroundings and from the lake surface. If the residual precipitation (227 mm) represents the amount of water from the surroundings and the mean discharge from the total area is 35.93 m³/sec (mean discharge at Motala in the period 1921–50, according to Melin 1955), the following relation is obtained:

$(P - E)$ of Lake Vättern $= 65$ mm.

With 554 mm of precipitation (Table 4), the evaporation from the lake surface should be 489 mm. In comparison with the results of investigations on the evaporation of some smaller lakes in southern central Sweden (A. Wallén 1927, Melin 1928, H. Johnsson 1946), this estimation can be judged to be roughly correct.

Winds

Introduction

The wind-generated sea is the most active and specific agency of the shore processes. The wind is also of importance as the force involved in wind set-up, water circulation, ice drift and eolic processes.

By methods of wave forecasting, the wave characteristics can be computed, when the wind velocity within the fetch length and the wind duration are known. The wind conditions can be worked out from synoptic charts or from local records. The last-mentioned method is often the only possible one in enclosed or semi-enclosed bodies of water (Beach Erosion Board 1961, p. 14). If the aim is to estimate the conditions which govern the trend of morphological development, wind statistics for many years are needed. Unfortunately, the wind variables involved are often incompletely known, because of insufficient instrumental equipment at the meteorological stations.

In Sweden, most stations estimate wind direction and wind velocity (wind force according to the Beaufort scale) only ocularly at certain fixed times during the day. At other stations, the two characteristics are read from an indicating wind vane and an anemometer. Some stations have a continuously recording velocity meter (anemometer), and very few have instruments recording both velocity and direction.

Wind-velocity and wind-frequency distributions

The wind-velocity distribution in relation to time is made up of a series of wind situations, interrupted by periods of calm or almost calm weather. The types of distributions of wind situations depend on the general atmospheric circulation and local influences, and are therefore different in different climatic regions. The cyclonic activity in northern Europe and the balance between marine and continental air masses at the margin of the continent cause wind situations of different genesis and of generally short duration in Sweden.

The wind-frequency distributions, obtained from stations which make observations at fixed intervals, do not refer to the individual wind situations, and in areas of unsteady conditions they therefore characterize the existing winds especially incompletely.

An analysis of wind statistics from 19 Swedish reporting stations, 15 situated on the western coast and on the Baltic, and 4 on Lake Vänern and Lake Vättern, has shown that the wind frequency distributions, all directions included, were of the same type (Norrman 1959). The frequencies of the distributions computed at the Meteorological and Hydrological Office were based on a 10-year period (1939–48) with an average number of 12,000 observations from each station. The wind velocity was represented by the wind force according to the Beaufort scale.

It was found that the frequencies were not far from being normally distributed, when the Beaufort units were taken as a linear scale, but the curves generally had some positive skewness. The following relations were obtained between the means of the actual distributions and the means (m) of the fitted normal distributions,

$$m = -4.28 + 3.95\ \overline{X}^{0.5}. \tag{4}$$

The actual means ranged from 2.4 to 4.8 Beaufort (B). The deviations from this relationship were very small, only one station having a mean (m) deviating more than 0.06 Beaufort units. The standard deviations of the frequency distributions ranged from 1.47 to 1.76 Beaufort units for the inland stations and from 1.53 to 2.20 for the coastal stations.

In order to estimate the frequencies of the rare high velocities, the statistical material obtained from a period of some years is often looked upon as a sample, and the expected frequencies for an infinitely long time are calculated from formulas of the following type (the constants may vary):

$$F_i = \frac{3n_i - 1}{3N + 1}\,100, \tag{5}$$

where F_i is the expected frequency of velocities $\leqq i$, n_i is the number of observations with a velocity $\leqq i$, and N is the total number of observations. The constants of this formula have been suggested by Samsioe (1961). From the expected frequency distribution the frequencies of higher velocities than those observed are estimated by extrapolation. The danger of such extrapolations is that, although higher velocities may occur, they are not necessarily correlated to the known distribution. This may, for example, be the case when the higher velocities are the result of weather situations of a type not apparent in the sample.

Estimated and true wind

When the wind velocities are estimated by observations of the wind effect, according to the Beaufort scale, there are naturally possibilities of errors that are often difficult to estimate. If the observations are made by one person, the errors could at best be constant (cf. Östman 1928, p. 14).

By comparisons between estimated Beaufort values and wind velocities in m/sec, computed from recordings, it has been possible to check the frequency distribution of the Visingsö station. The anemometer is placed 6 m above the ground and is surrounded by open fields, except to the west, where there are some deciduous trees. The instrument is of the wind-way measuring type. As the records, belonging to the Meteorological and Hydrological Office, had never been used, the velocities had to be computed. This was done not only to check the estimated velocities, but primarily to obtain figures of wind duration. The mean velocity in every second half-hour in the period 1946–50 was computed.

The relationship between estimated wind in Beaufort units and recorded velocities in m/sec was found to be very constant, but not in full agreement with the official relationship used by the Office. In Fig. 14, the variations of the estimations, based on corresponding recorded velocities, are shown. The data are sampled from observations made at 4 or 6 a.m. during January and February 1946–50.

The observations are in good agreement with the equivalents in m/sec of the Beaufort scale up to about 5 Beaufort. For higher velocities the observer has followed a linear relationship and overestimated the velocities. From the frequency distributions of observations at different Beaufort units the estimations can be judged to have been consistent and quite constant.

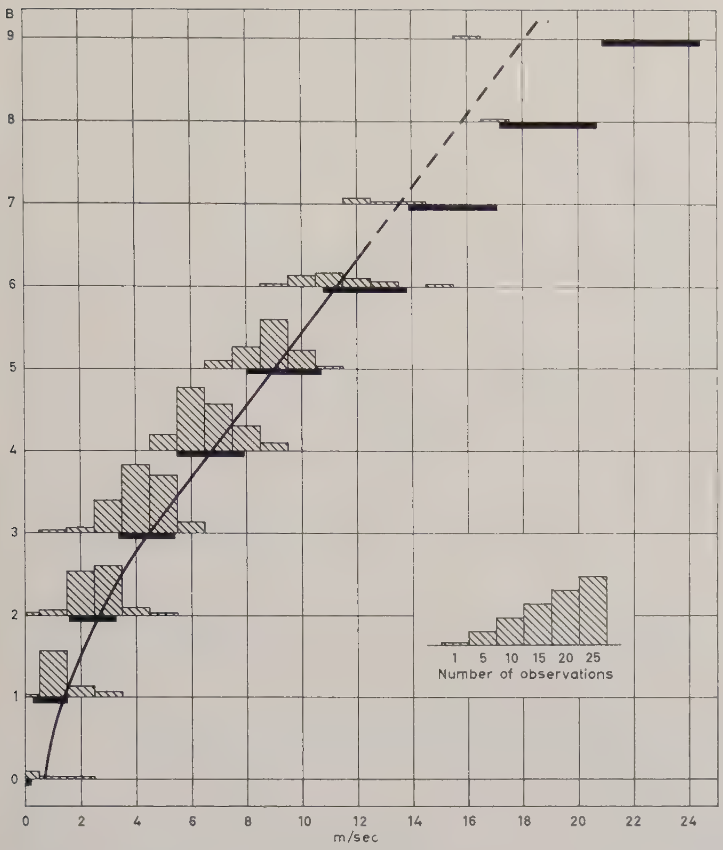


Fig. 14. Distributions of estimated wind velocities in Beaufort units (B) in relation to simultanously recorded mean velocities (0.5 hours) in m/sec. The heavy black lines are the official Beaufort intervals in m/sec, and the graph shows the relation between the estimated wind velocities and the recorded mean velocities.

The velocity distributions for every month and for the year, calculated from recorded data from the period 1946–50 (mean velocities of 41,616 half-hours), are illustrated in Fig. 15. The curves are drawn from calculated frequencies of velocities, exceeding 6, 9, 12, 15, and 18 m/sec. The frequency distribution of estimated velocities for the year is also shown.

The tendency of the curves to be straight lines in a normal probability scale is obvious. Only during November to April were velocities higher than 15 m/sec recorded, their frequency during the year being 0.25 %. From the estimated velocities this frequency should be nearly twice as high (0.45 %).

Wind velocities over land and water

The vertical wind-velocity profile depends on the roughness of the ground, gustiness of the wind, stability conditions, and the vertical temperature distribution. The velocity profiles, especially of grasslands, have been investigated in detail by Deacon (1953), who found that in the case of adiabatic temperature stratification conditions, the profile could be expressed by the logarithmic law in the form proposed by Nikuradse and Schlichting.

The roughness parameter of downland varies, according to Deacon (op. cit., Fig. 8), between 1.0 and 2.0 cm in winter, and between 2.0 and 4.0 cm in summer; for short grass it is 0.7 cm and for the sea with wind velocities > 7 m/sec. about 0.05 cm.

Over open water surfaces the wind will gain in velocity, and when measured at the downward end of the fetch, velocities higher than the average over the water will be obtained. The relation between wind velocity over land and water has been investigated in two inland reservoirs in the United States (Beach Erosion Board, 1962). A well-defined relationship could not be established, and quite large variations

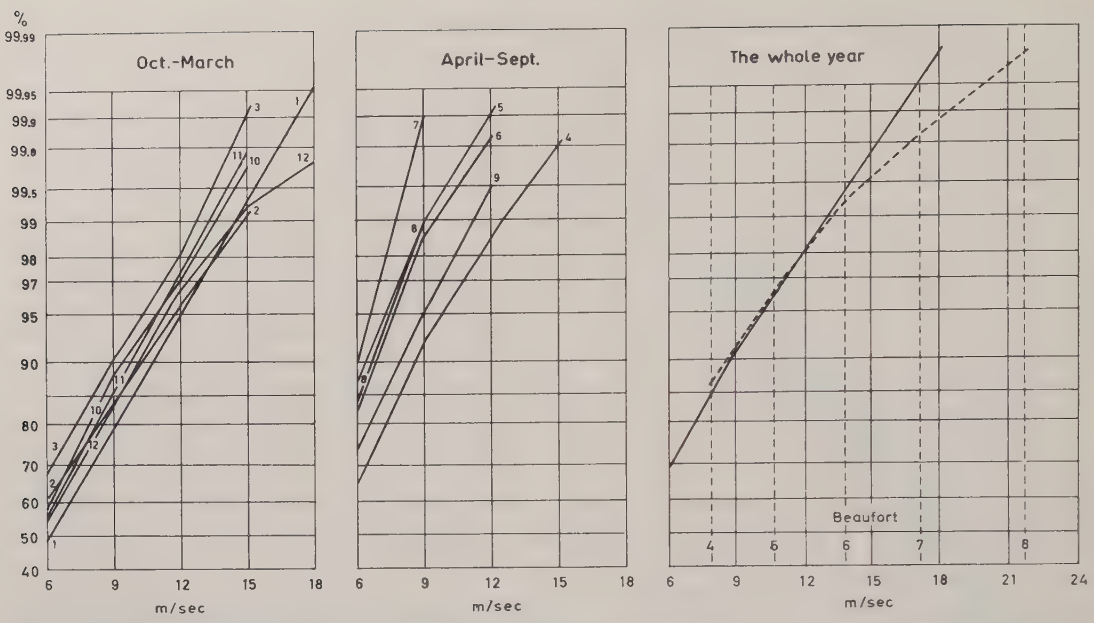


Fig. 15. Wind-frequency distribution of Visingsö (1946—50) for single months (1—12) and for the year, based on recorded velocities in m/sec and estimated velocities on the Beaufort scale (dotted line). Normal probability scale.

could also be expected, as variable conditions in the near surface layers of the atmosphere and variable roughness were not taken into account. However, a relationship based on median values has been suggested. The description of this relationship (Beach Erosion Board, 1962, Pl. 29) is somewhat ambiguous, but from a personal communication from the Beach Erosion Board (Thorndike Saville, Jr) it has been possible to draw the graphs in Fig. 16.

The graph q_A shows the ratio between the wind velocity at the downward (U_{Do}) and upward (U_{Up}) ends of the fetch. For fetches, longer than about 10 km, this ratio is constant (1.31). The graph q_B shows the ratio between the mean velocity over the fetch ($\bar{U}_W$) and the velocity at the upward end of the fetch. In the case of a fetch longer than 10 km, the mean velocity over the water is approximately

where the fetch (F) is measured in kilometres.

$$\bar{U}_W = (1.31 - 0.85/F)\,U_{Up}, \qquad (6)$$

If the wind is measured at the downward end of the fetch, the mean wind velocity over the water is obtained from

$$\bar{U}_W = (q_B/q_A)\,U_{Do}, \qquad (7)$$

which for fetches longer than 10 km can be written

$$\bar{U}_W = 1.31^{-1}\,(1.31 - 0.85/F)\,U_{Do}. \qquad (8)$$

As regards Lake Vättern, no fetches of importance are less than 5 km, and most fetches are longer than 10 km. From the graphs of Fig. 16 and the formula (7) we find that the average velocity over the water is 90–95 % of the velocity at the downward end of the fetch for fetches shorter than about 13 km; for longer fetches this average velocity is higher than 95 %.

In wave measurements carried out in the Huskvarna bay (Samuelsson, Svensson, and Teofilusson 1960) relations deviating from Bretschneider's wave-forecast diagrams (rev. ed. 1954) were obtained. These deviations were ascribed to the wind velocities taken from the Jönköping meteorological station, situated at an airfield 2000 m south of the shore, and with the anemometer placed 16 m above the ground.

The wind velocities at the shore and at the station were synchronously measured (during four days with quite weak winds, cf. Fig. 33), and the velocities at the shore (with onshore

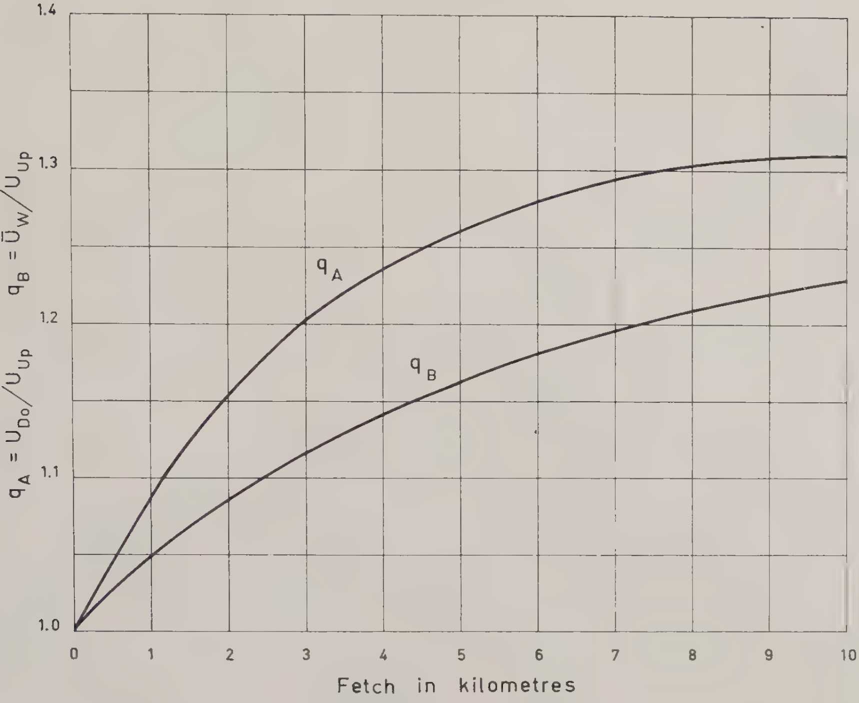


Fig. 16. Relationships between wind velocity at the downward and upward end of the fetch (q_A), and between mean wind velocity over the fetch and the velocity at the upward end (q_B). (based on Beach Erosion Board 1962, Pl. 29).

winds) were found to be about 1.6 times higher at the same height (Hedar 1960 *a*).

This constant ratio has been used (Hedar, op. cit.) in order to correct the wind statistics of the Jönköping station, which must be erroneous, because, on account of the unequal roughness parameters, the ratio will decrease with height. In order to reduce the velocities at 16 m to velocities at 8 m (the height used in wave forecasting), tre ratio 1.6 has been multiplied by $2^{-0.25}$ (= 0.84). Presuming a logarithmic wind profile, this constant corresponds to a roughness parameter of about 20 cm, which seems very unlikely.

However, the velocities at the Jönköping station are certainly lower than the velocities at the shore. If the roughness parameter at the shore is 0.05 cm, the roughness parameter of ground at the airfield around the station 1.0 cm, the profiles logarithmic near the ground and the ratio 1.6 at a height of about 3 m, then the ratio at a height of 16 m is about 1.38. At the shore, the velocity at 8 m is 93 % of the velocity at 16 m, and thus the velocities measured at the station should be multiplied by 1.28 in order to obtain the velocities at 8 m at the end of the fetches.

If the general ratio 1.31 in Fig. 16 (referring to a height of 8 m) is applied, with the same assumptions as regards roughness, the factor transforming the measurements at 16 m to shore velocities at 8 m will be 1.26.

At Jönköping the lower velocity at the meteorological station is due to roughness elements of a permanent character—buildings at the shore front, low trees and bushes. At Visingsö the anemometer is placed 6 m above the ground and about 12 m above the lake level. The open fields around the station have a roughness varying with the seasons. During late autumn and winter, when the wind velocities are high, the snow-covered fields have a very low roughness.

The frequency distributions of wind velocities of winds blowing along the lake have been calculated by formula (5), using statistics from Jönköping (1939–48) and Visingsö (1946–50). The total number of reported velocities

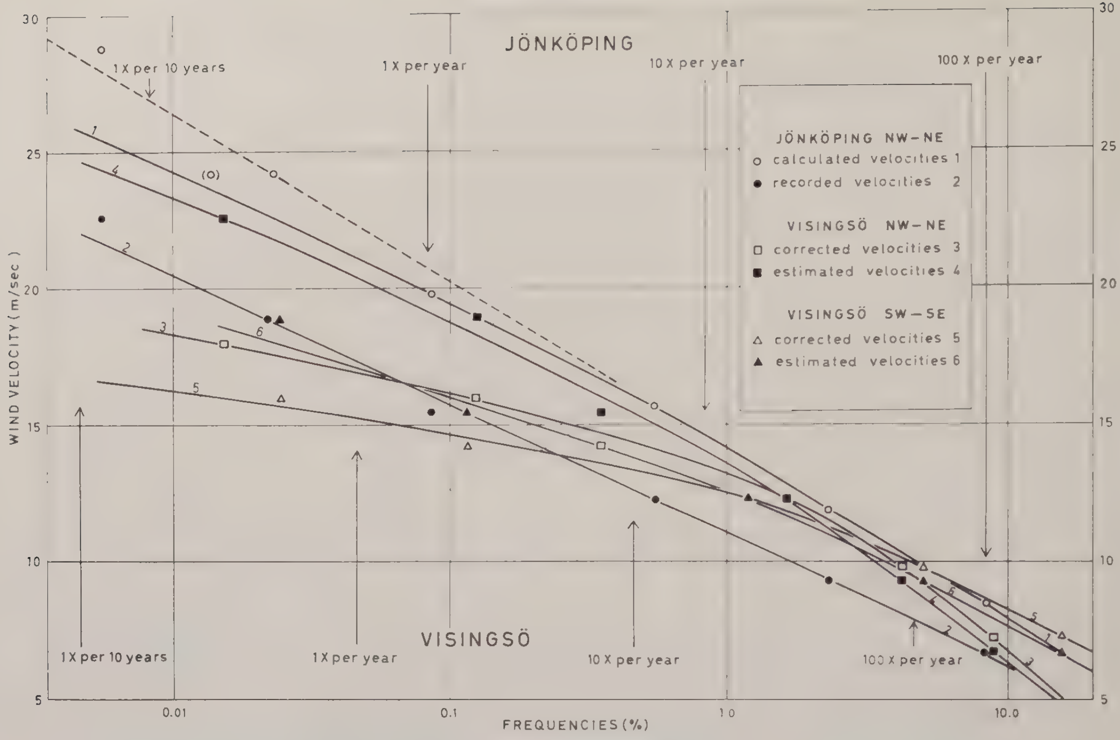


Fig. 17. Frequency of winds exceeding different wind velocities (cf. the text).

from Jönköping is 12,034 and from Visingsö 10,956. The azimuthal intervals NW.–NE. at Jönköping and NW.–NE. and SW.–SE. at Visingsö will now be compared. It can be expected that the frequency distributions of the northern intervals will show about the same distributions.

In Fig. 17 the distribution of recorded velocities (≧ 4 *B*) at Jönköping is shown, together with the distribution calculated with the ratio 1.28 between winds at the station and winds at the shore. The distributions at Visingsö are represented by the estimated reported velocities and velocities corrected by comparison with recorded velocities of the same period.

We find that in the northern interval the calculated Jönköping distribution is about the same as the Visingsö distribution based on reported velocities. This is especially the case if one single velocity of 9 Beaufort reported at Jönköping is excluded. If it is assumed that the Jönköping distribution shows the correct velocities at the height of 8 m at the shore, then the velocities reported from Visingsö can also be judged to represent the velocities at the shores. That is to say, the overestimation of the velocities at a height of 6 m at the station roughly corresponds to a decrease in velocity over the fields and the difference in velocity between 6 and 8 m.

The true velocity distributions over the water at the end of the fetches can only be found by a special extensive investigation covering all different weather situations and stability conditions, but, as far as comparative studies of exposure are concerned, it seems that the distributions based on the statistics from Visingsö are sufficiently correct.

Azimuthal wind-frequency distributions in the Vättern basin

The only central place where wind observations are made is on Visingsö island. Because of the bordering hills the distributions obtained at stations on the mainland can be expected to be greatly affected by these hills. The distributions based on all wind forces and on winds equal to and stronger than 4 Beaufort at Jönköping, Visingsö and Karlsborg are

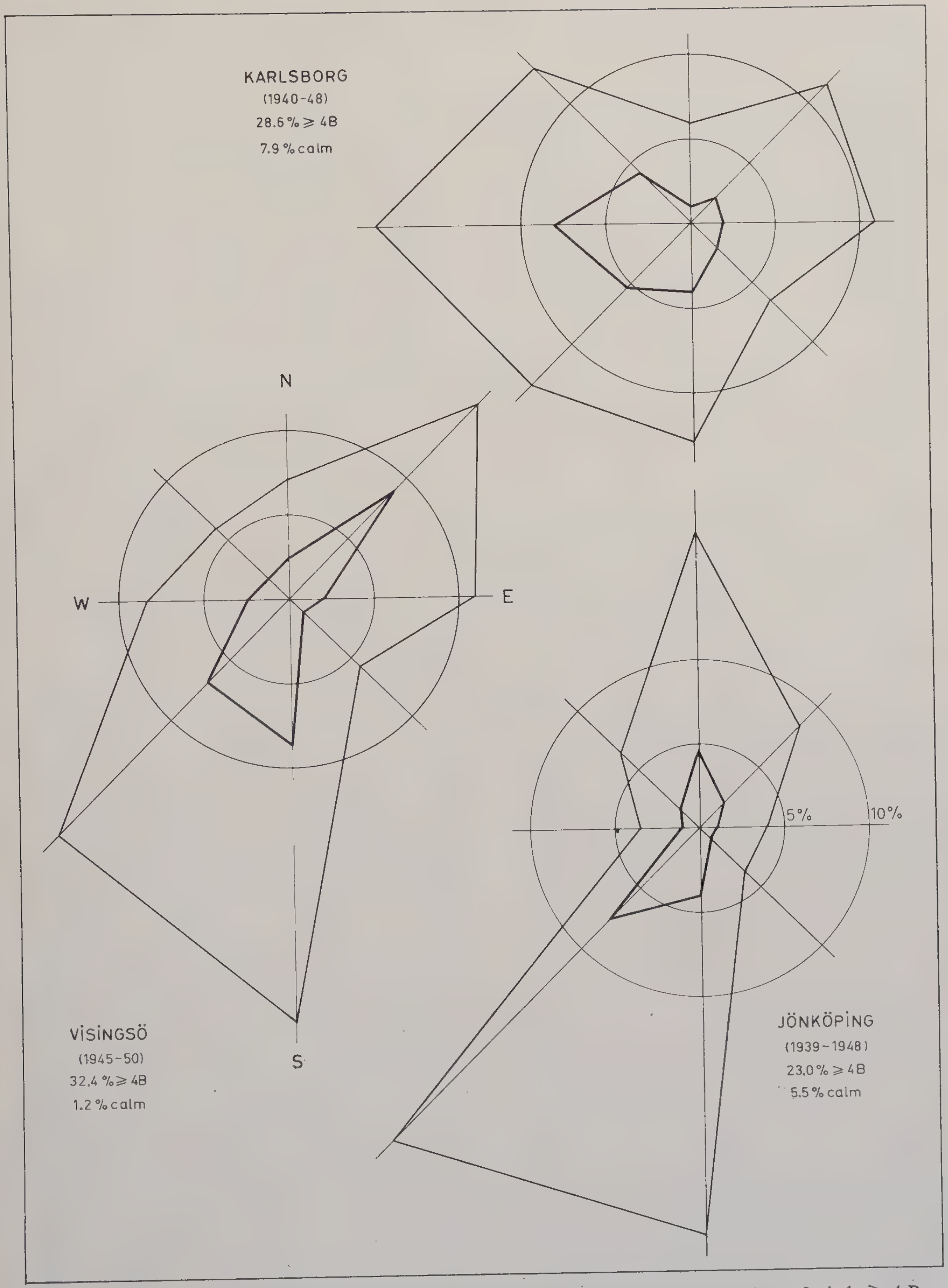


Fig. 18. Wind-frequency distributions for Karlsborg, Visingsö and Jönköping. The distributions of winds ≧ 4 Beaufort marked by heavy line.

shown in Fig. 18. The diagrams have been calculated from the reported winds without corrections.

The effect of the topography of the basin is obvious. The majority of the winds blowing past Visingsö from the north-east quadrant are forced to follow the direction of the steep hillsides of southern Lake Vättern, and they approach Jönköping from a more northerly direction. In the same way winds from the south-east in Jönköping tend to approach Visingsö from a more southerly direction. It can also be seen that Jönköping, situated in a rather narrow depression, is more sheltered from easterly and westerly winds than Visingsö. The directions predominant when the frequencies are counted on all winds are still more predominant when only the stronger winds are taken into account.

The wind distribution of Karlsborg differs from the other two by its maximum in the NW.–SW. quadrant. This maximum is caused by the low passes in the border north-west and south-west of Karlsborg. As at Visingsö, there is a secondary maximum in the north-east quadrant.

The conditions in the Karlsborg area are exceptional and in general it can be estimated that most of the winds blow in directions fairly close to the orientation of the basin. The diagrams show that the western shore is exposed to more winds from the north-east quadrant than from the south-east, but, because of its NNW.–SSE. orientation, it is also affected by southerly winds. The eastern shore is exposed to more winds from the south-west quadrant than from the north-east, but northerly winds also affect this shore.

Distributions of wind durations at Visingsö

As previously mentioned, the mean wind velocity in every second half-hour from the records of the Visingsö station in the period 1946–50 has been calculated. In order to examine the distributions of wind durations, capable of raising a sea with important effects on the shores, a concept called "wind occasion" has been introduced. From field studies it was found that with winds of 6 m/sec or more there was generally a sea that caused sand to move along the shores. This wind velocity is lower than that used by Schou in his formula on wind work (1945), where only winds stronger than 4 B (> 8 m/sec) are taken into account.

A "wind occasion" has been defined as a situation in which at least two consecutive half-hours have a velocity $\geqq 6$ m/sec. However, in a series an isolated half-hour may have a lower velocity without breaking the occasion. As only every second half-hour has been calculated, the minimum duration of a wind occasion is 2.0 ± 0.5 hours. From the fact that, during the wind occasions studied, the velocity very seldom dropped below 6 m/sec and that there were few single half-hours with this velocity, it can be judged that the concept represents definite wind situations.

The frequency distributions of the duration of wind occasions are illustrated in Fig. 19. The scale is log-normal. The seasonal variation is obvious, and the difference is large between summer with mostly short occasions (50 % less than 4 hours and 90 % less than 10 hours), and winter with high frequencies of rather long duration (50 % less than about 12 hours and 90 % less than about 50 hours). The curves of spring and autumn lie quite close together in between these extremes. The longer durations in the autumn mainly appear in November.

The graph shows that there is a seasonal variation in wave attack on the shores, not only from variation in wind velocity but also from the duration, as there will be a longer time for the waves to grow during late autumn and winter than during spring and summer.

It is of interrest to know whether there is a relationship between the duration of wind occasions and the frequencies of higher velocities. However, the velocity distribution during a wind occasion is not always as simple as a stage with a gradually growing wind, a peak, and a stage of dying out. From the point of view of time there are also flat distributions and distributions with several maxima. It is therefore difficult to give a statistical description of the wind occasions. Although it covers 5 years, the material is not large enough for long-term estimates of the time distribution of extreme velocities.

In Fig. 20, the relation between the duration of wind occasions and the occurrence of a velocity $\geqq 9$ m/sec (mean velocity of at least one half-hour) is plotted. The curves connect mean values, calculated on the wind occasions grouped in 5-hour intervals. The curves can be

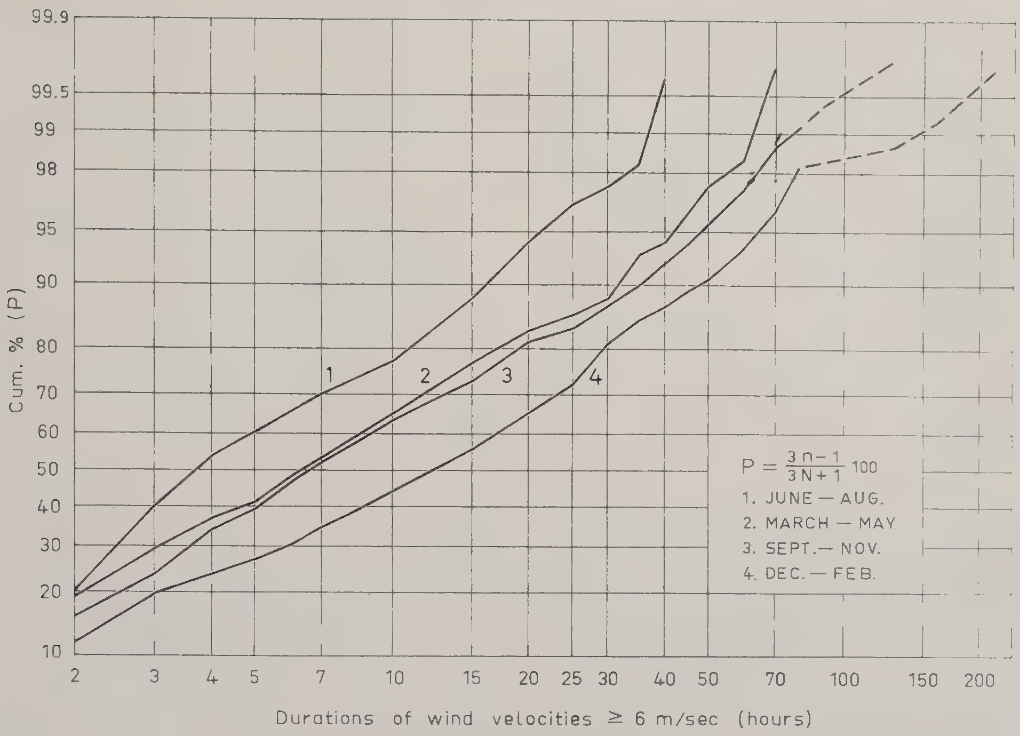


Fig. 19. Cumulative distribution of wind occasions with a wind velocity continuously exceeding 6 m/sec. Normal probability scale. Based on wind statistics from Visingsö for the period 1945–50.

looked upon as probability graphs. We find that the probability for a 9 m/sec wind is very small for durations less than 5 hours. For all seasons, except summer, the curves lie quite close together up to 60 % or 15 hours of duration. At durations longer than 10–15 hours 50 % of the occasions involve the occurrence of a 9 m/sec wind. In summertime a duration of 15–20 hours is needed for the same percentage. The summer and winter curves are most regular, with a steady increase in probability of higher velocities with increasing durations. The spring curve reflects the unsteady conditions of this season. For a long

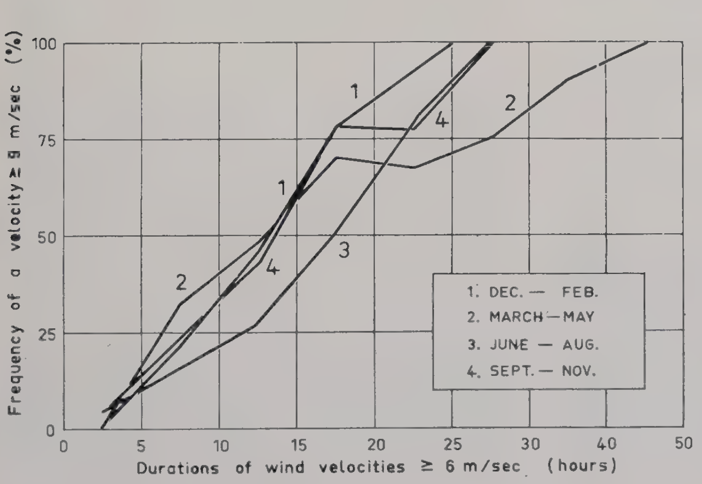


Fig. 20. Frequencies of a wind velocity ≥ 9 m/sec in relation to duration of winds ≥ 6 m/sec. Based on wind statistics from Visingsö for the period 1945–60.

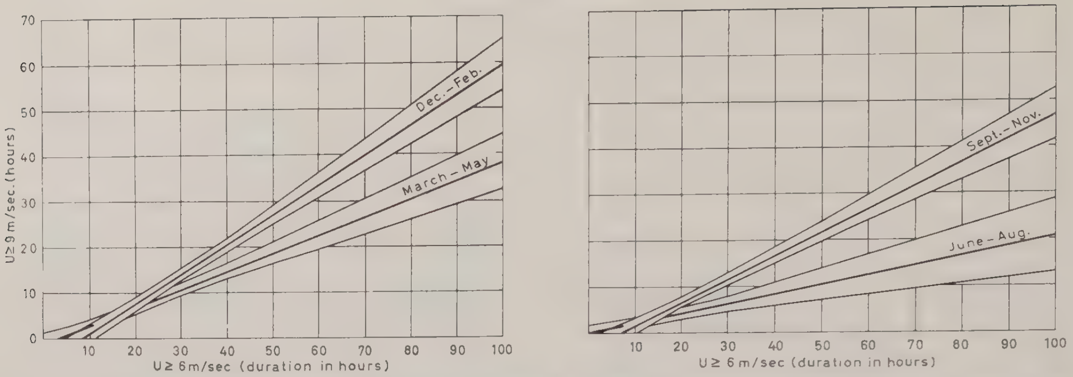


Fig. 21. Relationships between duration of wind velocities ≥ 6 m/sec and time of velocities ≥ 9 m/sec for the four seasons. 95 % confidence limits for estimated relationship. Based on wind statistics from Visingsö for the period 1945–50.

time, the mean wind velocity can oscillate between 6 and 9 m/sec without rising to higher values. The standard error of the curves is naturally large. In the central part of the curves it is about 10–15 %, being smallest in winter and autumn, and with the longer durations of the spring higher than 20 % (Norrman 1959).

The relation between the duration of wind occasions (U ≧ 6 m/sec) and the total time of velocities ≧ 9 m/sec has been calculated. This relation is only based on occasions with a velocity ≧ 9 m/sec. The scatter diagrams did not indicate a calculation of any other relationship than a linear regression could give. The regression lines of the four seasons are shown in Fig. 21. With the longer durations in winter, velocities ≧ 9 m/sec exist during an average of 50 % of the time, in autumn during 40–50 %, in spring during slightly less than 40 %, and in summer during about 20 %. The comparatively high frequencies of stronger winds with short durations in summer (cf. Fig. 21) can be judged to reflect the short but rather violent thunder-storms occasionally appearing in July and early August.

The limits of the estimations of the relationships for the five-year period as a whole indicate that, for single events, the variations are large, but they also show statistically well-defined differences between the seasons.

The frequency of wind velocities ≧ 12 m/sec is low (cf. Fig. 15). During the five-year period such a velocity was only recorded once in summertime. Although based on a small material, statistical characteristics of the distribution during wind occasions are given in Table 5.

Table 5. Distribution in hours of wind velocities ≧ 12 m/sec during wind occasions.

	Q_1	Md	Q_3	P_{90}
Winter	2	5	13	31
Spring	1	2	7	8
Autumn	1	4	7	16

It should be remembered that the figures in Table 5 show the total duration of the higher velocities, and accordingly, their continuous durations are still shorter.

Hitherto, only durations have been dealt with irrespective of wind directions. Where there was continuous recording of velocities, the wind direction was only read off six times a day between 5 a.m. and 7 p.m. This makes a proper analysis of the relation between directions and durations impossible. However, it is essential to know, whether there are large differences or not in this relation for different directions, and because of the shape of Lake Vättern, as well as the azimuthal distributions of wind frequency, it is most important to know about differences between winds blowing in opposite directions along the lake. In order to examine this problem from the material available, the number of azimuthal observations within the intervals NNE.–NE. and S.–WSW. has been counted for every single wind occasion, and these numbers have been placed in relation to the total number of ob-

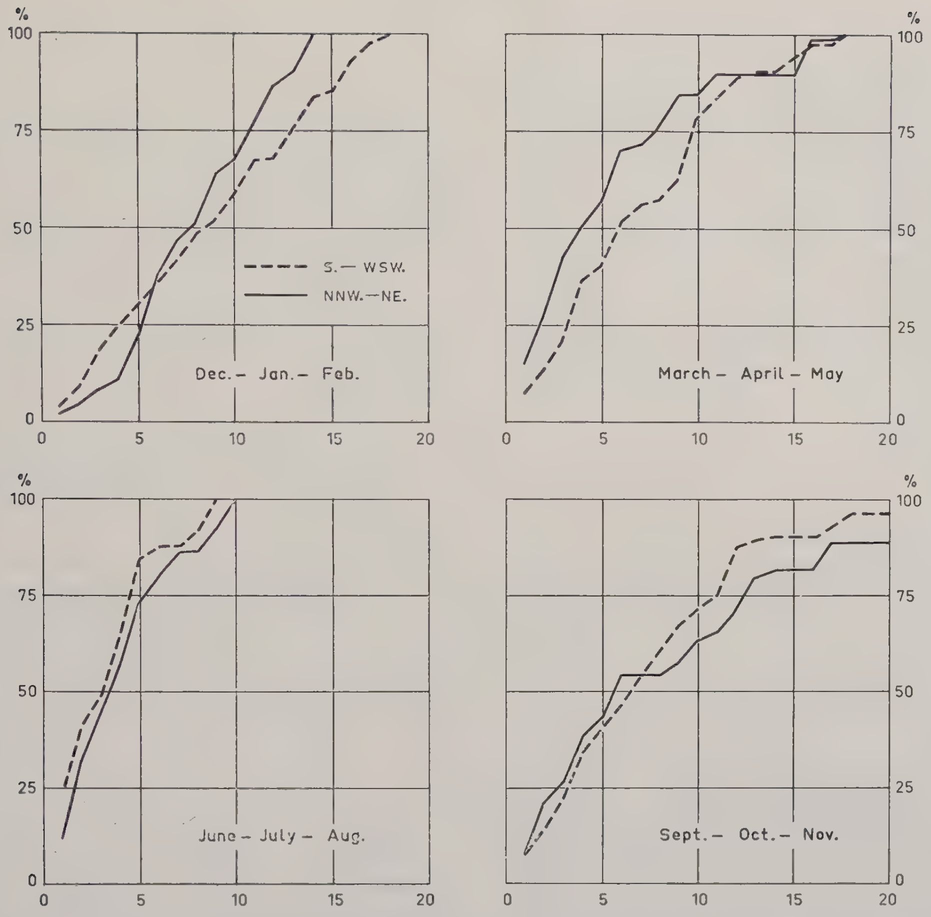


Fig. 22. Percentual distribution of wind observations ≧ 6 m/sec in the intervals S.–WSW. and NNW. –NE. in relation to the total number of observations of the associated wind durations with a velocity ≧ 6 m/sec. Based on wind statistics from Visingsö for the period 1945–50.

servations of the occasion. The percentual distributions have then been calculated (Fig. 22).

The graphs of Fig. 22 thus indicate to what degree the winds of the two azimuthal intervals are associated with different durations of wind occasions. The seasonal variation follows from the relations shown in Fig. 19, but the curves also show that there is no significant difference between the two azimuthal intervals. That is to say, as far as the 6 m/sec level is concerned, the wave attack from northerly and southerly winds is not significantly unequal because of different distributions of duration. Such a homogeneity with respect to the distribution of durations is of importance, when the relative effect of wave attack from different directions is estimated from wind-frequency diagrams, as, for example, in the formulas of resultant wind work.

Hydrology

Introduction

The first works dealing with the hydrology and hydrography of Lake Vättern (Hjärne 1706 (Swedish translation 1916), Block 1708, Tiselius 1723, 1730) dwell upon the "peculiarities" of the lake, such as mirages, presumed subterranean connections with distant lakes (Lake Geneva, the Black Sea), etc. They have been commented by Nerman (1898).

Several of the investigations carried out on the lake have been concerned with zoological or botanical problems. Among the works which have some hydrological aspects or are mainly on its hydrology, the following should be mentioned: Pettersson (1902), Grenander (1904), Ekman (1914), Björkman (1917–21), Halbfass (1923), F. Bergsten (1926), Stålberg (1939) and Melin (1955).

A manuscript on the stage regulation of Lake Vättern and the discharge from the Lake, compiled by S. Ernfors (1961), formerly the Senior State Hydrologist, has been made available to me. The following description of the effect of artificial measures on the discharge from the lake is to a large extent based on this manuscript.

Discharge and run-off

The discharge and run-off from Lake Vättern have the characteristics, tabulated below, which are based on data from the period 1858–1939 (Melin 1955, p. 217).

Table 6. Discharge and specific run-off in the period 1858-1939.

	Max.		Mean			Min.	
	Abs.	Mean	Abs.	Mean	Min.	Mean	Min.
m^3/sec	100	57	74.7	42.0	14.3	28	8.9
1/sec km^2	16	9.0	11.7	6.6	2.2	4.4	1.4

	Duration for a year and for the period						
	50 %			75 %			95%
	Abs.	Min.	Per.	Abs.	Min.	Per.	Per.
m^3/sec	79	15.3	41	63	12.7	32	20
1/sec km^2	12.4	2.4	6.5	9.9	2.0	5.0	3.2

The natural discharge is through the Motala river to the Baltic. Before the Motala power station was built (1919–21), the discharge was affected by water mills and fisheries in the stream and by some drainage through the eastern part of the Göta Canal, connecting the lake with the Baltic. The construction of the canal began in 1810, and was finished in 1832. When the first flour-mills were erected is unknown, but there were undoubtedly mills in medieval times. The milling trade ended in 1921.

According to an investigation carried out for the Water Court (the Court's decision of Aug. 16th, 1926), because of the possibility of changing the elevations of the mill gates, the discharge could be varied between 7 and 8 m^3/sec at lower water stages, between 12 and 13 m^3/sec at higher stages, and possibly between 20 and 30 m^3/sec at the highest water stages. However, according to Ernfors (1961, p. 1), the real variation has probably never at any water stage exceeded 6 to 8 m^3/sec, though it may have reached 10 m^3/sec a few times. His conclusions are based on the material available at the Meteorological and Hydrological Office, and he declares that the theoretically highest possible damming was never used at high water stages.

The amounts necessary for the locks in the Göta Canal are small, but the canal was also used for drainage through the Motala Works until 1921. The Works were founded for engineering purposes connected with the construction of the canal. The total drainage in this way can, according to a memorandum from the State Water Board (A. Löfroth, March 3rd, 1921), be estimated at a maximum of about 9 m^3/sec. During the latter part of the 19th century, it is supposed to have been less than 4 m^3/sec. Hence, this drainage counterbalanced the damming effect of constructions in the natural stream.

According to the official discharge curve (established in the ordinance dated Dec. 19th, 1928), a change of 1 m^3/sec in discharge corresponds to a change in water stage of about 3 cm at the lowest water stages (about 88.0 m above sea-level), about 1 cm at 88.6 m, and about 0.8 cm at the highest water stages. The theoretically highest possible damming effect from the works in the stream (30 m^3/sec) could thus produce a rise in the water level of about 24 cm over the highest natural level.

In some documents from the 17th century, extensive damage in Jönköping is reported and is claimed by the citizens of this town to have been caused by the damming of the water by mills and fisheries in the Motala river (Björkman 1919, pp. 328–332). That there was damage because of wave erosion and flooding is certain, but this could also have happened without artificial measures.

Water stage

Observations

Prior to the Swedish precision levelling of 1900, the water stage of Lake Vättern could only be very approximately related to the sea-level. When the Göta Canal was constructed, the first excavation was found to be 0.6 m too shallow in relation to the determined sailing depth, because of an erroneous estimation of the lowest water stage of the lake (Nerman 1898, p. 224). From the time of the construction and onwards, water-stage observations have been carried out in Motala. The observations since 1832 have been preserved. Observations from the winter months are sporadic before 1858.

Until 1929 the water stage was indicated in relation to the lock threshold at Motala. Since then it has been given in metres above sea-level, the limnograph at Motala (67–154 *Motala*) being connected to the precision levelling of 1900 by the levelling point 67–2938 *Motala a*. Since 1900 the water stage has also been read or recorded by the municipal authorities of Jönköping.

Water-stage regulation

Since 1928 the water stage has been regulated by the ordinances of 1926, 1938, 1940 and 1958. In the current ordinance, that of 1958, a continuous year's regulation was permitted. The permission includes the right to exceed the natural water stage by 9 cm, when this stage is 88.30 m or lower. The right of damming is gradually reduced, and is 0 cm at a water stage of 88.70 m or higher. The stage may also be lowered to 88.30 m or to the natural water stage, if this is lower than 88.30 m, but with some modifications and restrictions at low water stages, involving, amongst other things, certain cm/day calculations at water stages between 88.41 to 88.30 m. The regulation is managed by the State water works in Motala, and inspected by the Meteorological and Hydrological Office.

Water-stage variations

The yearly water stage generally varies in a certain direction over several years. This course is exemplified by the yearly and monthly means of the period 1901–30 (Fig. 23). During the year the water stage is as a rule highest in late summer and lowest in wintertime.

The percentual distribution of the water stages, based on the monthly means of the non-regulated period 1858–1927, is illustrated in Fig. 25. Between the levels 88.2 to 88.8 m, the water stage is almost normally distributed. It varies within this interval for 93 % of the time. The median value (88.50 m) is identical with the mean level.

The distribution during the period of stage regulation (1928–61) is also shown in Fig. 25. In both periods, the water stage has varied within the interval 88.00–88.52 m for 55 % of the time. The curve of the regulated period, however, lies on an average 2–3 cm lower. In the interval 88.52–88.76 m (40 % of the time) the curves coincide, and during the regulated period a monthly mean exceeding 88.88 m has not been recorded. During the preceding period higher stages are recorded for almost 2 % of the time, which would correspond to 8 months of the regulated period 1928–61. Summing up, the differences in percentual distribution between the non-regulated and the regulated periods concerned are small.

The very small variations in the water stages of Lake Vättern become apparent, when compared with those of Lake Vänern (A. Wallén 1910), the adjacent and largest lake in Sweden. In Fig. 25 the median of the water-stage distribution of Lake Vänern has been placed at the same level as that of Lake Vättern.

The monthly means of the regulated (actual) water stage and the calculated (natural) water stage in the period 1947–61 have been placed together in Fig. 24. The figure shows that the regulation has mainly had a raising influence on the water stage, both at high and at low water stages. An exception is the year 1959, and the first six months of 1960, when the regulated surface lay below the calculated, natural surface at all water stages. The curves also show that the influence of the regulation has been small. The maximum rise for any month is 5 cm, and the regulated level is on an average almost 1 cm higher than the calculated, natural one.

The distribution of the mean duration of periods, continuously exceeding different levels, has been calculated for every 5th cm on the basis of the monthly means of 1858–1927 (Fig. 26). The maximum and minimum values of the

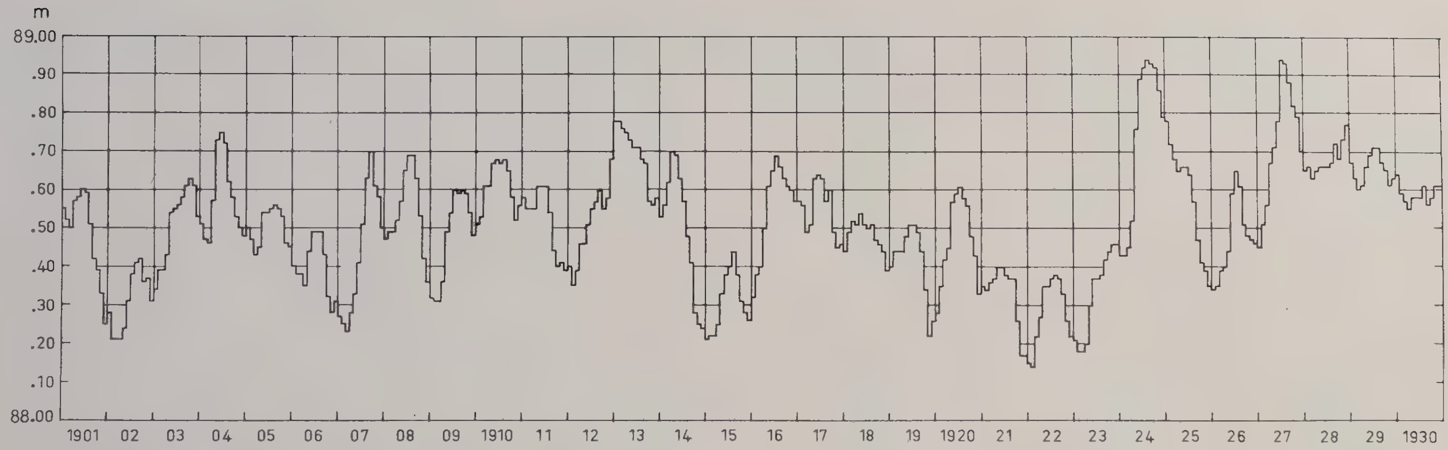


Fig. 23. Water stage of Lake Vättern. Monthly means 1901–30.

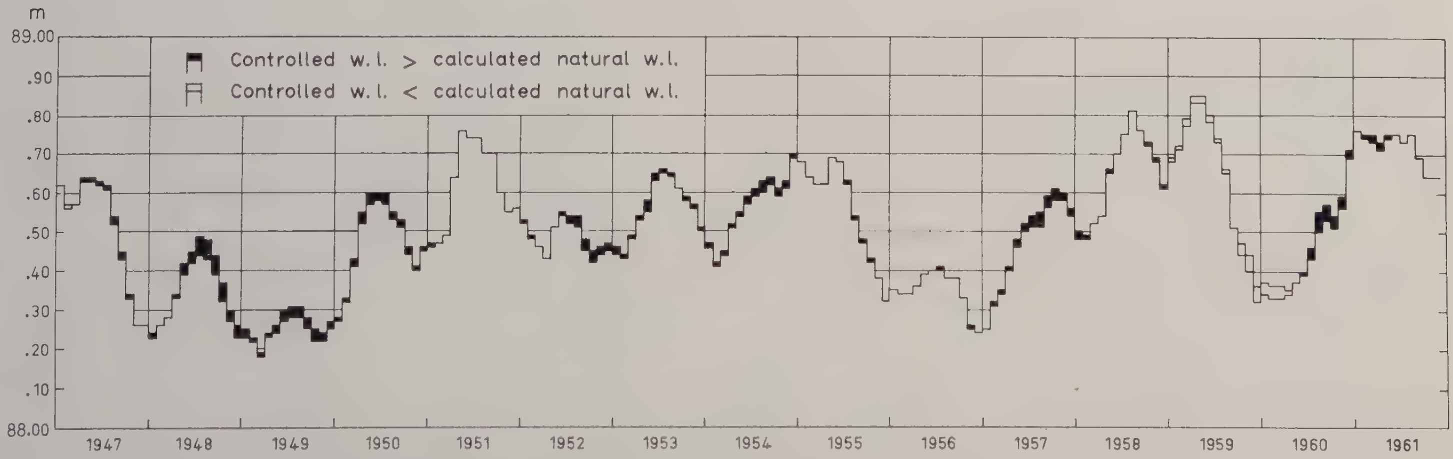


Fig. 24. Water stage of Lake Vättern 1947–61 and the effect of water-stage control.

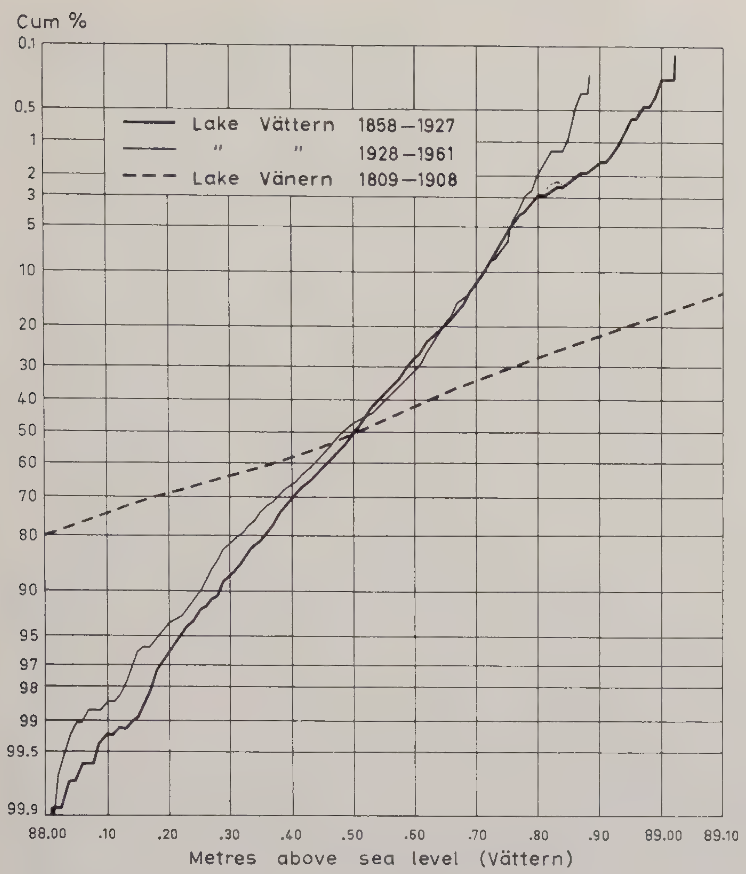


Fig. 25. Water-stage distribution of Lake Vättern and Lake Vänern. Normal probability scale.

period are 88.00 and 89.02 m. A smoothed curve, based on the means of the interval 88.25–88.75 m, has been calculated by the method of least squares, applied to the function

$$\log t = a + b \log h, \qquad (9)$$

where t is the mean duration in months and h the level in dm above 88.00 m.

The following relationship was obtained:

$$\log t = 2.650 - 2.376 \log h. \qquad (10)$$

The curve (Fig. 26) fits the actual values well within the interval 88.15–88.75 m. The deviation of the lowest values is due to two separate occasions, when the water stage was extremely low, in January 1859, when the level was 88.00 m, and in January and February 1888, when the level was 88.08. On no other occasion during the last 70 years, has the monthly mean fallen below 88.15. The lowest values, connected by the ocularly fitted, dotted curve in Fig. 26, can be regarded as anomalous. The deviations in the upper part of the curve are also due to a few exceptional measurements. The level 88.95 has been exceeded once.

The frequency distribution of periods exceeding different levels is also shown in Fig. 26. It can be approximately fitted to a normal distribution, but the actual distribution is somewhat skewed towards higher levels. The quite regular variations in water stage are apparent from the curves of duration and frequency.

Thermal conditions

Although several series of measurements of the water temperature have been taken ever since the 18th century (for a bibliography, cf. Stålberg 1939), no general, regional survey has as yet been carried out.

The lake is known to be very cold, a fact quite well known to me from the Scuba dives.

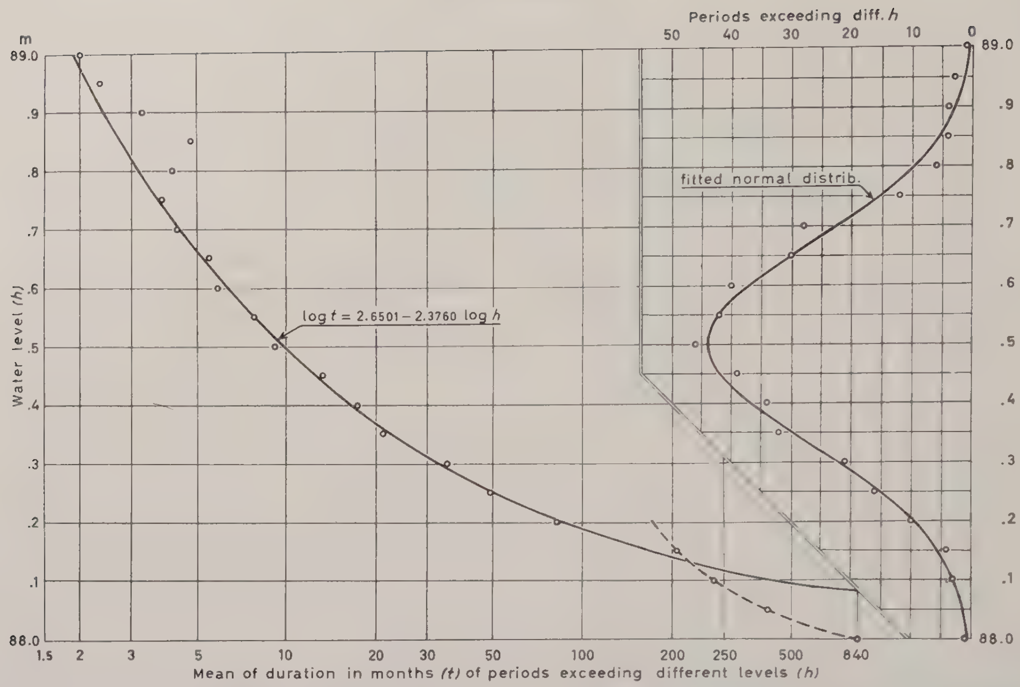


Fig. 26. Water-stage distribution with respect to the duration of periods exceeding different levels estimated on the monthly means of the years 1858–1927 (840 months).

The lowest temperature reported at the bottom, + 0.38° C, was measured by L. Larsson in March 1900 (O. Pettersson 1902) at a depth of 95 m.

In winter, the water of the basin is almost homothermal (Stålberg 1939, p. 13), which can be ascribed to the late freezing over or total lack of covering ice (cf. Fig. 28).

In summer, the thermocline is generally vaguely accentuated, and embraces a con-

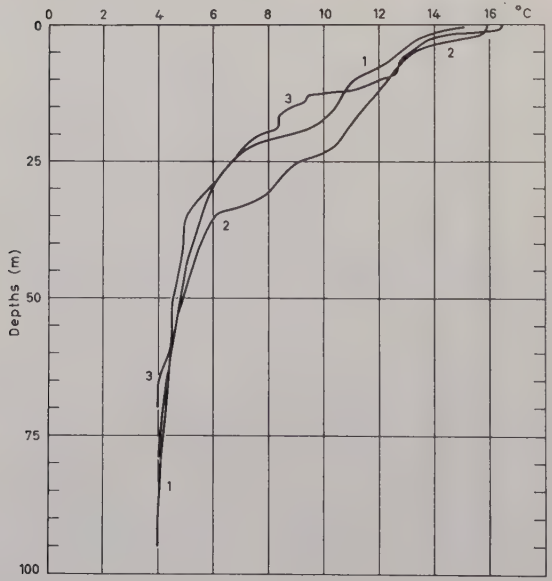


Fig. 27. Water temperature on July 17th, 1962 (1) halfway between Jönköping and Visingsö island, (2) in the deep channel between Gränna and Visingsö island, (3) in waters between Karlsborg and Fjuk island. Thermistor recordings by the State Water Inspectorate.

siderable layer of water (Stålberg 1939, p. 13). The conditions are illustrated by some series of measurements made in July 1962 by the State Water Inspectorate (Fig. 27). The curves reflect the mixing of warmed surface water by wave activity.

The effect of wind drift is sometimes easily recognized by accumulation of quite warm surface water (about 15–18° C) in shallow windward areas and the welling up of water with a temperature around 6° C in leeward areas, the optimal effect being obtained by a steady, not too strong breeze, stronger winds having a mixing effect on account of the wave motion. To judge from notes from the diving operations, the depth of this warm, poorly mixed, surface water does not exceed about 2 m.

The onshore drift of warm surface water may be of some importance as regards the distribution of turbid water from the breaker zone and the deposition of the suspended material.

Transparency

Lake Vättern is not only known as a cold but also as a very clear lake. The high transparency reflects the oligotroph character of the lake. The shores and the bottoms are very sterile, except for some inner parts of the Motala bay and sheltered areas in the archipelago. Ekman (1915) and Thunmark (1937) have reported a visual depth of 17.1–17.7 m in the open parts of the lake. In the summer of 1962, the Sechi disc was found to disappear at a depth of about 13 m in the deep waters around Visingsö. It is uncertain if this difference lies within the normal variations or if it is the result of a growing pollution from outlets.

Some daylight penetrates to large depths. In a search for a lost bottom sampler, the diver was able to inspect the bottom sediments when swimming 1 m above the bottom at a depth of 38 m off Hjo. Figs. 137–138 are reproduced from colour slides taken without flash-light at a depth of 17 m.

There is a high percentage of oxygen even near the bottom in the deep troughs (Ekman 1915), and the proportion of deposited organic material is generally very small, except in the central parts of the deep troughs, where the probability of sedimentation of abraded, inorganic material is extremely low.

Ice conditions

All the small lakes of southern central Sweden are generally covered with ice and snow for several months in the winters (cf. J. V. Eriksson 1920). In contrast, Lake Vättern is only occasionally frozen over from shore to shore. Stålberg (1939, p. 14) estimates the frequency of this occurring as once in ten years. The causes of this open water can be traced to the large mass of water to be cooled, the exposed lake surface, and the rather strong winds in winter, which impede the formation of a stabilized, thermal stratification and which can also break up thin sheets of newly formed ice.

The meteorological stations around Lake Vättern have not carried out properly organized ice observations, but in their logbooks notes on the ice conditions can be found. The severe winters of 1940–42 prompted some work on ice conditions in Sweden, and from data available at the Meteorological and Hydrological Office a record of the ice conditions in Lake Vättern for the period 1880/81–1939/40 was compiled by Liljequist (1941). From this paper, further reports from the Office, notes made at the harbour office on Visingsö, and from my own observations, the conditions during the period 1880/81–1960/61 have been worked out as shown in Fig. 28. From the reports and notes, it is often difficult to state the exact area which has been covered by ice, and also the exact data concerning the freezing over and breaking up. However, these possible errors are of minor importance from a geomorphological point of view.

The late freezing and late breaking up are typical. The month having the highest frequency of ice cover is March, but December, January and February have a lower mean temperature, and before the breaking up the deciduous trees are often already green. This means that there is a long period in late autumn and winter with open water and frozen ground and a period in spring with lake ice but ice-free beaches and slopes.

Small lakes may be covered almost instantly with a thin sheet of ice, or the ice rim gradually advances outwards from the shores. In the most sheltered parts of the archipelago of northern Lake Vättern, the same pattern is found, but not in the open parts of the lake. There, ice layers first start to grow in shallow

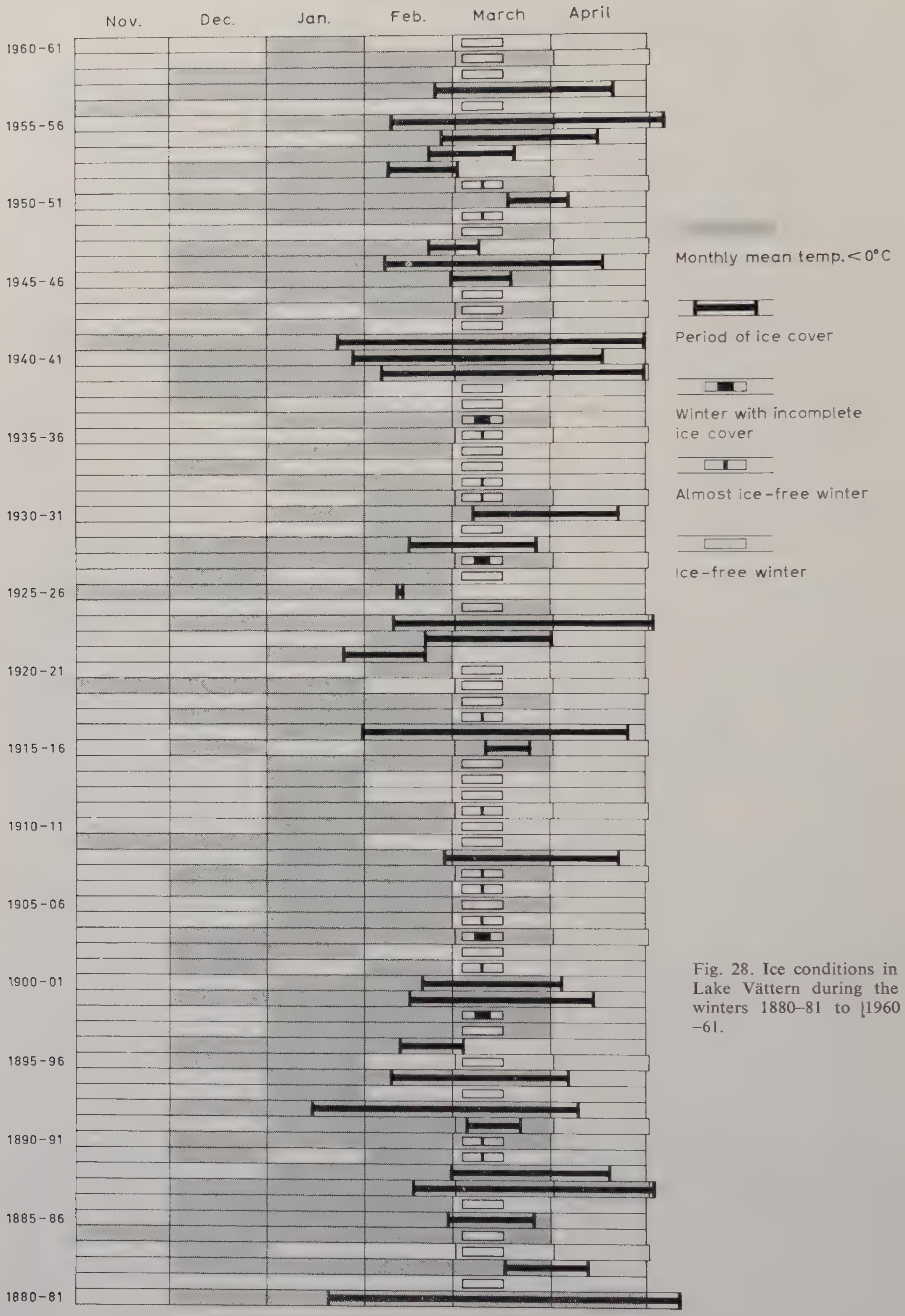


Fig. 28. Ice conditions in Lake Vättern during the winters 1880–81 to 1960–61.

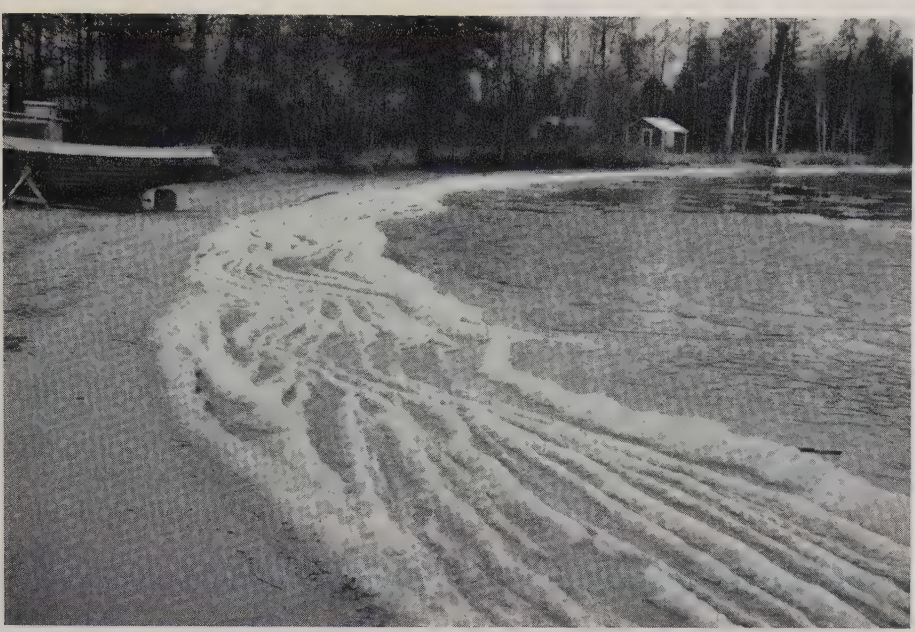


Fig. 29. Ridges of frazil ice and ice floes drifted into the bay of the outer end of Enebågen 11 km south of Karlsborg. Nov. 1963.

bays, but this ice is almost always broken up by wind and waves. The broken-up pieces drifting around become rounded by rubbing, and between them the water is filled with frazil ice (Fig. 29).

When the drift is hard, broken ice is driven ashore, and the pieces are piled up in ice walls. Then the shores are in general already glaciated by spray from the surf and swash. Sometimes, the ice walls are forced up on the beaches, but usually there is a nearshore belt of broken ice, frozen down to the bottom, and then the walls are piled up at the front of this ice bench at a water depth of about 0.5–1.0 m.

During the winter of 1955–56, the ice around the island of Visingsö was especially studied (Norrman 1959). A broad, concentric system of ice walls was developed along the northern and western shores, with flat belts of re-frozen, broken ice between these ridges, reflecting the variations between periods of moderate and hard weather. The ice was thinnest in the walls lying nearest the shore. The width of the system (up to about 300 m) was undoubtedly related to the existence of a shallow bottom. Drillings showed that even in the outermost walls, where the water depth was 1.5–2.0 m, ice-floes were pressed down to the bottom (Fig. 30).

With time larger fields of broken ice freeze, and when the lake is at last frozen over, one can see how the cover, far from the shores, is made up of a mosaic pattern of ice-floes. In less severe winters, the formation of lake ice can stop at any stage, and if the winter is mild, there are only glaciated shores and some fringes of floating ice in sheltered areas.

Thermal contraction and extension of the ice cover cause large cracks, generally running diagonally across the lake. Several cracks are found in the same position from year to year (cf. Cholnoky 1908, Pl. XIV). One, for example, runs from some hundred metres south of Gränna harbour towards the north-eastern spit of Visingsö.

Reliable information about ice thickness is scanty, but for a total ice cover the figures normally lie in the interval 0.2–0.6 m. There is generally not very much snow on the ice, dry snow being blown ashore.

In spring, the ice is fretted by sunshine and rain. It melts around dark obstacles, such as stones and breakwaters, and along the shores, especially the stony cliff shores. In this way,

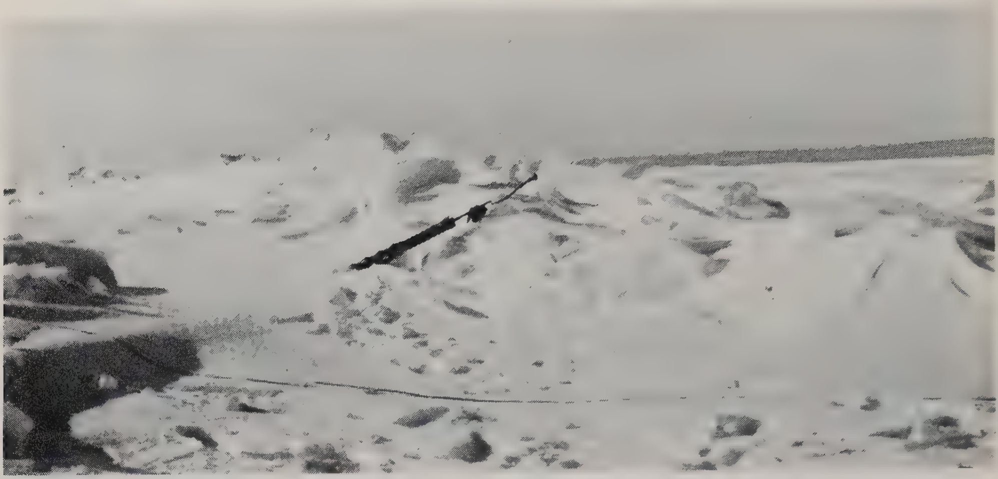


Fig. 30. Thin floes of new ice piled up in ridges. In the foreground an ice-core drill. Feb. 1956.

channels of open water can be formed along some shores exposed to insolation.

The fretted, floating ice becomes porous with a vertical pipiness and, although still thick, its strength is then very much reduced, especially as regards bendings, perpendicular to the surface. If a block of ice is cut out, it is found to be composed of quite loosely united, vertical sticks or needles. When the ice has become very pipy, before the cover is broken up by a strong wind, these needles rapidly fall apart, and the whole lake can be ice-free in a few days. If there are fields of stronger ice, they will drift, and thick ice walls may be pressed up on the shores (Fig. 31).

The relation between ice conditions and morphological processes is treated in a special chapter.

Wind drift and water-level undulations

Water-level variations within the lake are developed by wind drift and differences in barometric pressure over the surface. The free undulations of the lake have been investigated by F. Bergsten (1926), who analysed them by means of Crystal's method, and compared the

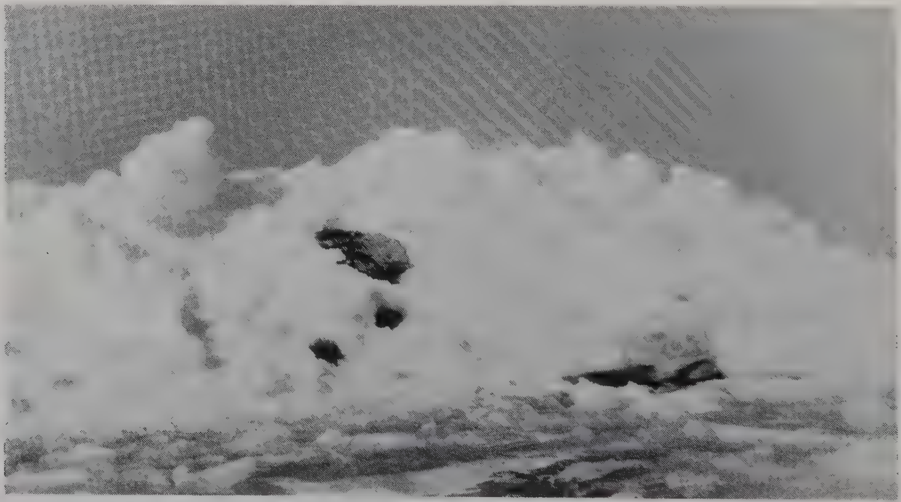


Fig. 31. Ice rampart pushed up by wind drift in spring. The porous ice cracks in blocks. Height of rampart 4 m. April 1956.

theoretically calculated characteristics with records from limnographs. He was able to determine the position of the nodes of uninodal, binodal and trinodal longitudinal seiches, and also found undulations of higher orders. The period of the uninodal wave was almost 3 hours (177.94 minutes), and the maximum amplitude recorded was 12–13 cm. The longitudinal binodal seiches had a maximum amplitude of about 8 cm. However, the amplitudes are generally much lower than these maxima. "During at least 75 % of the time, the size of the amplitude is only one or two centimetres" (information from Bergsten, cited by Stålberg 1939, p. 12).

Bergsten also investigated the origin of the seiches, and found the wind effect to be of outstanding importance. In his calculations of wind effect he made use of Hayford's proposal that the wind force (kg/m²) is a function of $U^{2.4}$, where U is the wind velocity. The constant of the function needed was empirically determined to be valid for Lake Vättern. Bergsten mentions the possibility of the applied exponent of U being too large, but, because of lack of good wind data, this could not be determined (1926, p. 50). The water body was represented by 27 defined cross-sections (1926, Fig. 1), drawn perpendicular to the axis of the lake, but some sections were added in the area of complicated bottom topography at the northern end of the lake. The wind effect for different wind directions is only given explicitly for his limnograph station at Bastedalen, situated at the entrance to the northern archipelago, 119 km from Jönköping. Although surface profiles could be constructed from Bergsten's data, a new approach has been made, in order to try a first approximation with fewer cross-sections and a better verified exponent of U. In this attempt the method proposed by Hellström (1941, 1953) has been applied.

The differential equation for the gradient of the water surface is expressed by

$$dz_0/dx = k_1\tau_s/\varrho g(z_0 - z_1), \qquad (11)$$

where z_0 is the coordinate of the water surface, z_1 the coordinate of the bottom, τ_s the wind force or tangential stress on the water surface, ϱ the mass density of the water an k_1 a constant (1.5).

The relation between wind force and wind velocity is given by the equation

$$\tau_s = k_2U^{1.8}, \qquad (12)$$

where U is the wind velocity in m/sec and k_2 a constant (0.00040, Hellström 1953), and the wind force is expressed in kg/m².

For practical purposes the depth $(z_0 - z_1)$ can be represented by the mean depths of sections perpendicular to the wind direction. The surface slope is calculated by successive integration of the gradients of the sections, and for a closed water body the level of the surface is determined on the assumption of a constant water mass.

Only the longitudinal surface slopes, corresponding to winds from the ENE. and WSW., have been calculated. The calculations were based on 15 equally spaced cross-sections, and the mean depths determined from the isobaths in Pl. 1. The bottom profile of mean depth and the surface profile for a wind velocity of 12 m/sec blowing from the ENE. are shown in Fig. 32. The surface profile is uncertain north of the northernmost trough because of the irregular topography. Calculated changes in level at Jönköping and differences between Jönköping and a line at the latitude of Motala (*J–M*) for different wind velocities are tabulated.

Table 7. Effect of wind on the water level in Jönköping, and the difference in level between Jönköping and Motala

U m/sec	Jönk. cm	(*J—M*) cm
6	2.6	3.6
12	8.9	12.4
18	18.5	25.7
24	31.0	43.2

Although the values at Bastedalen are somewhat uncertain, they are of interest because a direct comparison with Bergsten's figures (1926, Table 5) is possible (Table 8).

Owing to the higher exponent of U applied by Bergsten, the increase in wind effect with wind velocity is more rapid in his calculations. The too high exponent is to a certain degree compensated by the empirically determined constant. The two functions compared give the same figures for a wind velocity around 10 m/sec, and as can be found from the table,

Table 8. Levels at Bastedalen according to calculations by F. Bergsten (1962) and by Hellström's formula (1963).

Wind velocity		Wind drift (cm)	
B[1]	m/sec	1926	1963
4	7	2.0	3.0
6	11	5.9	6.8
9	18.5	20.6	18.5
10	22	31.3	25.0

[1] The relation between Beaufort scale (B) and wind velocity in m/sec is according to Bergsten, and differs from that otherwise used in this paper.

the differences are small for wind velocities less than 20 m/sec.

It is also of interest to note the quite small difference in the position of the nodal line. Bergsten gives the figure 69.2 km from Jönköping, and in Fig. 32 it is situated 72.5 km from the same place.

The most remarkable, longitudinal, wind drift mentioned by Bergsten occurred in 1917, when a sudden rise of 22 cm was recorded at Bastedalen, and about the same fall in Jönköping. A comparison between the mean wind velocity needed for such a wind effect and the wind records (cf. Fig. 17) also makes it probable that much higher figures can be expected extremely seldom.

With winds transverse to the axis of the lake the wind effect is generally insignificant, except for the Motala bay, where the lake is much broader than elsewhere and also shallow. At Motala, Bergsten recorded transversal seiches with a maximum amplitude reaching 20 cm, thus surpassing those of the longitudinal undulations (1926, p. 41). Hence, hydrographic levelling, based on the official limnograph situated at Motala, must be done with great care.

The effect on the shore morphology of the undulations giving larger water-level amplitudes towards the ends of the lake and in the Motala bay is difficult to estimate, as, because of the fetch length, the waves are also largest in these areas. However, the wind effect should be kept in mind, when constructions connected with the shores are projected.

Some authors (Ekman 1914, Lundqvist 1942 *a*) have worked on the currents, which have erosional effects on the bottom sediments of the lake without making any calculations or direct measurements. I have therefore estimated the current velocities that can be expected from wind drift. The estimation is based on the assumption that the highest velocities will be generated, if the tangential stress of a wind blowing along the lake suddenly ceases, and that the water mass raised on one side of the nodal line will pass this section in one-quarter of the period of the uninodal seiches.

If the period is 180 minutes, the section of the nodal line has an area of 0.67 km², and the water is raised, according to the figures given earlier, as a function of wind velocity, the mean current velocities (V_m), shown in Table 9, will be reached.

Table 9. Estimated mean current velocities in central Lake Vättern raised by wind drift.

U m/sec	V_m m/sec
6	0.006
12	0.021
18	0.044
24	0.073

As the flow cannot be concentrated to layers near the bottom by density stratification or in other ways, it seems impossible that currents with velocities exceeding the critical erosion velocities (cf. p. 82) could be generated in the open lake. As regards narrow channels, the conditions are different. I have observed a current velocity of about 1.5 m/sec in the small channel (about 4 m²) between the northern and southern parts of the island of Stora Röknen, the current going westwards, generated by a moderate wind from the west.

Waves

Methods of estimation

The relations between generating conditions and wave dimensions and methods of wave forecast in general are discussed in the first part of the chapter on shore processes.

It has been discussed if the wave forecasts of Sverdrup and Munk, Bretschneider, etc. (cf. p. 70), which are primarily based on oceanic conditions, can be used in Swedish coastal and inland waters (Hedar 1960*a*, Davidsson 1963).

When studying the abrasion on Visingsö is-

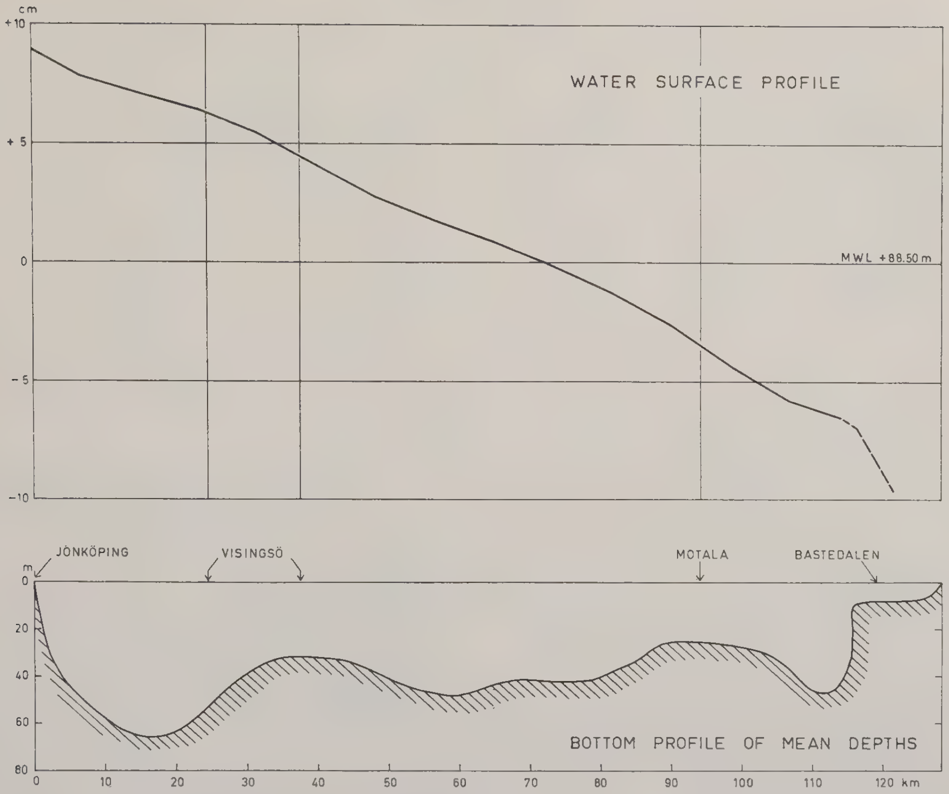


Fig. 32. The effect on the water surface of Lake Vättern of a 12 m/sec wind blowing along the lake from the north.

land in 1957, I set up an offshore wave pole with a float and carried out cinegraphic recordings. The pole was placed in a depth of 4 m about 150 m from the shore. Unfortunately the apparatus was soon afterwards demolished by a tanker running aground (a modified type of apparatus is described in Norrman 1963*b*).

In 1959, students from the Chalmers Institute of Technology, Göteborg, made wave measurements in the Huskvarna bay (Samuelsson, Svensson, and Teofilusson 1960). Their paper has not been printed, but the results of their investigations and of measurements in the Gulf of Bothnia have been summarized by Hedar (1960*a*). The measurements were ocularly made, one wave pole being anchored in deep water, and another standing in shallow water. Series of 250–300 consecutive waves were measured (height and period), and from these the statistical wave characteristics were calculated. The problem of getting adequate wind data has already been mentioned in the chapter on the winds (p. 45). Hedar states that, when the wind was measured on the beach at the end of the fetch, the results were in good agreement with Bretschneider's forecast diagram of 1954. However, some further comments on this subject are necessary.

The measurements for wind correction were carried out during 4 days (April 21st to 24th, 1960). I have examined the wind statistics for these days, recorded at Jönköping, Visingsö and Karlsborg. The maximum wind velocity (13.5 m/sec) was reported from Karlsborg. The velocities, reported at 0700, 1300 and 1900, are illustrated in Fig. 33. The great variation in direction and velocity of the synchronous observations is striking. It shows that the wind observed at one place often cannot be taken as representative of a large part of the lake. Some of the deviations can be explained by time lag, but the good agreement between wave forecast and measured data is still quite

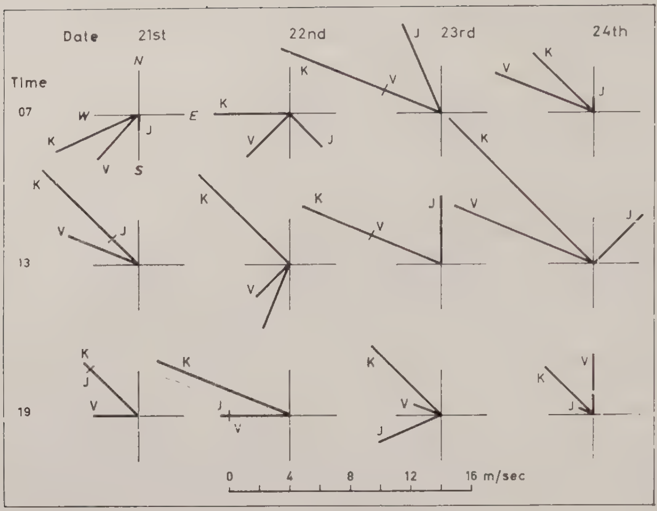


Fig. 33. Winds reported from Karlsborg (K), Visingsö (V), and Jönköpng (J) on the 21st to the 24th of April 1960.

astonishing. An explanation can be found from the fact that most of the waves were measured during weak wind conditions. In the original report there was only one occasion with a wind velocity exceeding 7 m/sec. As far as weak winds are concerned, only a small nearby wave generation area is of importance, and thus the local wind conditions could be correlated to the waves.

The investigation does not cover situations, in which the generation of a fully arisen sea is effectively limited by the fetch. If ordinary wave forecast diagrams are applied to strong winds blowing along the lake and the fetch is taken as the distance from shore to shore in the main direction of the wind, wave dimensions are obtained which are certainly not found in the lake. However, the results of wave measurements in inland reservoirs in the United States (Beach Erosion Board 1962) give a new possibility of wave forecast in lakes.

In order to produce wave data for Lake Vättern, effective fetches have been determined according to a method outlined by Saville (1954) and illustrated in the above-mentioned report from 1962. It is assumed that the wind has a generating effect within ± 45° of the main wind direction, that the fetch lengths (F_i) within this generation area are equal to the distance from shore to shore, and that the effective fetch (F_{eff}) can be expressed by

$$F_{eff} = \Sigma F_i \cos \alpha_i / \Sigma \cos \alpha_i, \qquad (13)$$

where α is the angle to the main wind direction, and the interval of α_i used is 6°, α_i ranging from ± (0°–42°).

By this method the effective fetch has been computed for 16 azimuthal directions (except at localities where some directions are sheltered by the mainland of the shore in question). An azimuthal distribution of effective fetches is thus obtained. Then the azimuthal distribution of significant wave heights for wind velocities of 6, 9, 12, 15, 18 and 21 m/sec has been determined from the diagram of Fig. 34. This diagram has been converted from Pl. 34 in the Beach Erosion Board's 1962 report. The azimuthal wave-height diagrams show the maximum significant heights corresponding to the effective fetch. The necessary durations to reach this condition are found in Fig. 34, in which also the relation between fetch and wave period is shown. The water depth is regarded as deep in relation to the waves generated.

A local wave-forecast diagram of this kind makes possible a direct comparison between waves generated from different directions and also comparisons between the potential exposure of different localities. The real quantitative distribution naturally depends on the wind distribution, and it cannot be found unless there is a continuous recording of wind velocity and direction. The problem of estimation of exposure from the incomplete wind statistics available in Lake Vättern is discussed in the chapter on the shore processes of the lake.

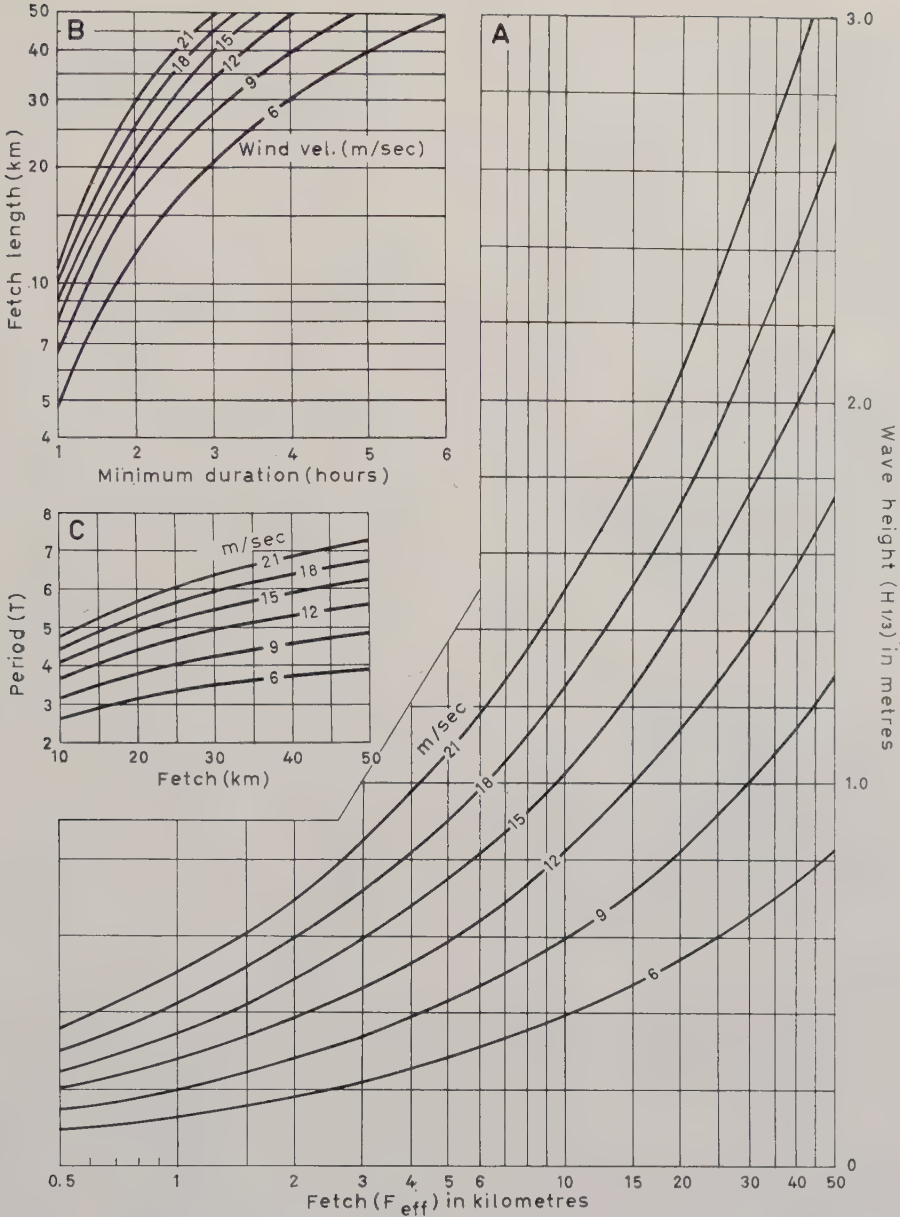


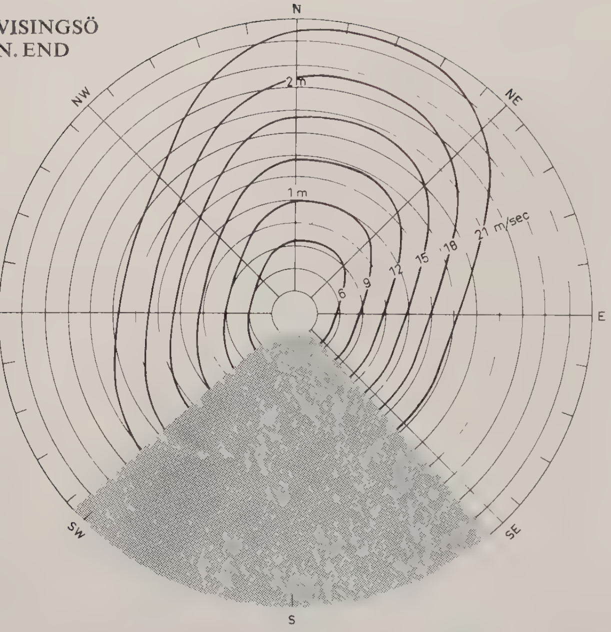
Fig. 34. *A*, significant wave heights for wind velocities between 6 and 21 m/sec and fetches between 0.5 and 50 km. *B*, relations between fetch length and minimum duration for different wind velocities. *C*, significant wave periods in relation to fetch and wind velocity (adapted from Beach Erosion Board 1961, 1962).

Local wave-forecast diagrams

In order to get wave characteristics for different parts of the lake, five localities were chosen: Jönköping bay, the southern end of Visingsö island, the northern end of Visingsö island, the Flisen shoal, and the Jungfrun skerry.

The diagram centre for the Jönköping and Visingsö localities was chosen offshore in deep

Fig. 35. Wave forecast diagrams, Significant wave heights for minimum durations corresponding to effective fetches.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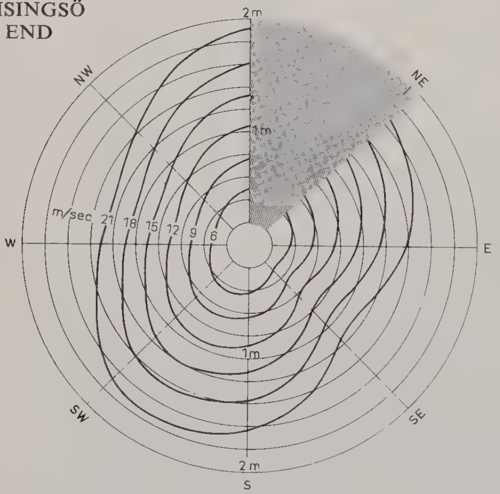
N. END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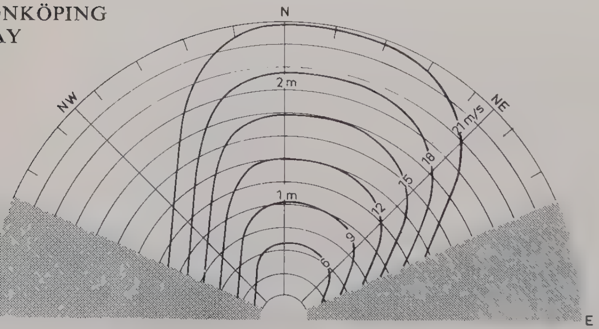
NKÖPING
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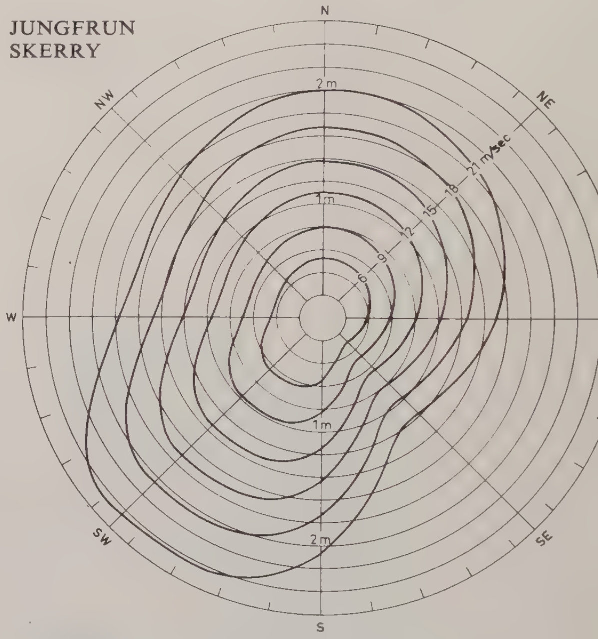
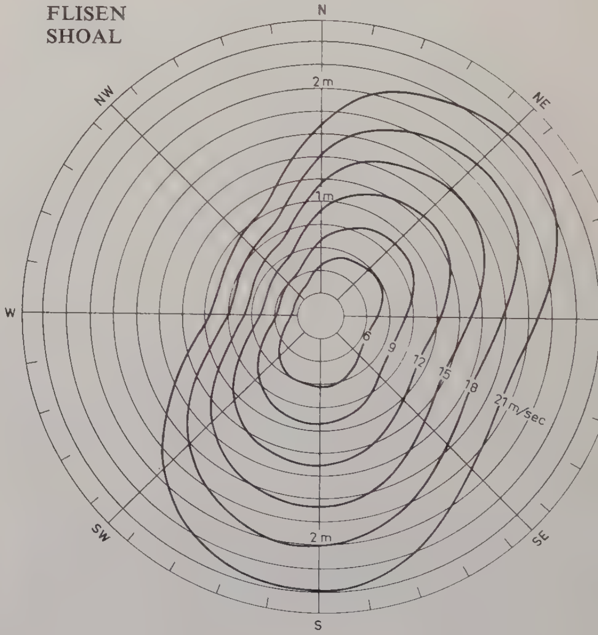

JUNGFRUN
SKERRY


FLISEN
SHOAL

water. As regards Jönköping, two alternative distributions are possible, one assuming that the fetches of direction N. 6° E. and N. 12° E. end at Visingsö island, and the other that there is no sheltering effect of this island at all. The sheltering effect is uncertain, but to judge from practical experience during several summers of navigation, this effect is limited to a narrow zone close to the island. The effective fetches of the Jungfrun skerry in directions between the north and north-east are probably overestimated, because there are skerries and shoals in this interval.

The local wave-forecast diagrams are shown in Fig. 35. The significant heights may look great but it should be remembered that they represent optimal conditions. As far as summer wind conditions are concerned, they have been qualitatively estimated as being of the correct order. Because of the limited fetches, the waves of the lake are steep and it is commonly known that special hulls are required for the safe navigation of small craft. The wave steepness for different wind velocities and fetches corresponding to the minimum duration are shown in the table below.

Table 10. Wave steepness corresponding to the minimum duration of the fetch.

U m/sec	Fetch (km)		
	10	20	30
6	0.037	0.035	0.035
9	0.038	0.037	0.036
12	0.039	0.038	0.037
15	0.040	0.039	0.038
18	0.041	0.040	0.039
21	0.042	0.041	0.040

The values of the table agree well with the results of the previously mentioned wave measurements in the Huskvarna bay. In these the wave steepness generally varied between 0.035 and 0.040.

The wave-forecast diagrams clearly demonstrate the great dependency on the fetch. Winds blowing along the lake generate waves, which are considerably larger than waves generated by winds blowing across it, and as the winds along the lake are most frequent, they will have a predominant effect on the shores.

The Flisen diagram is almost symmetrical with respect to waves generated along the lake–the "mean point" of these waves lies to the south of this shoal. As could be expected, the Visingsö diagrams are almost symmetrical in relation to the axis of the island. Because of its mid-lake position, the waves along the lake will predominate still more than on the mainland shores. It may be noted that, because of the configuration of the lake, the symmetry axis of the Jönköping diagram is directed towards the NNE.

PRINCIPLES OF SHORE PROCESSES AND SHORE MORPHOLOGY

Working agencies

Waves

Wave forecast

Wave data from different regions have been recorded and summarized in several works (Cornish 1904, Krümmel 1911, Bigelow and Edmondson 1947, Bruns 1955 etc.), but these observations were generally made without knowledge of or data on the agencies, which had developed the sea.

The relation between the wave characteristics and the factors, which determine the wave generation, could not be properly analysed until the method of hindcast was introduced. In hindcasts observed waves are compared with synoptic charts and recorded wind velocities. By quantitative analysis of the relation between the relevant factors of wave generation and the wave dimensions, empirical relationships have been established, which make possible the prediction of wave conditions from weather data (Sverdrup and Munk 1947). The methods of wave forecast have been continuously refined. Several somewhat different methods, based on wave spectra and the energy of wave spectra, have been proposed (Darbyshire 1952, 1955, Neumann 1953, 1955, Bretschneider 1952, 1954, 1958). Revised forecast diagrams for practical use, including diagrams for forecasting in shallow water areas (Thysse and Schief 1949, Bretschneider 1954), have been published by the Beach Erosion Board (1961).

The different methods of forecasting and the goodness of fit between observed data and forecasts have been discussed by King (1959). Wave variability, represented by the distributions of wave height and period from a large number of records, and wave length, computed from the period and utilized as $L = T^2$ (arbitrary units), have been studied by Bretschneider (1959). His investigations were carried out by means of marginal distributions and joint distributions in the technique of correlation analysis. Bretschneider analyses the theoretical differences between the wave spectra used by him, Darbyshire and Neumann, and also points out the effect of local conditions (bottom friction and currents of shallow water) on the quantitative results. In his latest revision of his forecasting relationships, the variation in spectral width is taken into account. The forecast diagrams published by the Beach Erosion Board in 1961 are based on the Sverdrup-Munk method, revised in 1961 by Bretschneider.

The applicability of wave-forecast diagrams to Swedish coastal and inland waters has been examined by Hedar (1960*a*) and Davidsson (1963). Hedar compared waves measured at Seskarö in the Gulf of Bothnia and in Lake Vättern with Bretschneider's diagram of 1954. He found no significant deviations. The wave measurements in Lake Vättern have already been discussed in the chapter on the hydrology. Davidsson (op. cit., p. 22), who made his measurements on the coast of Scania in southern Sweden, found that the diagram-determined wave height (Bretschneider 1958) was greater than the measured height. He mentions the probability of the influence of shallow water, but it also seems probable that too few waves were measured in each run to obtain a reliable result.

Wave generation in limited water areas

The size and energy of wind-generated waves grow with the wind velocity, the horizontal distance of open water over which the wind blows, and the duration during which the wind blows in essentially the same direction over the generating area–the fetch. But for wind velocity, fetch length and duration, the waves may also be dependent on the water depth and the width of the fetch, determined by the boundaries of the water surface.

For a given wind velocity, steady-state wave conditions develop over an increasing distance with increase in duration. A particular fetch length thus requires a minimum duration, dependent on the wind velocity, to reach this state. A certain duration correspondingly requires a minimum fetch length.

Within the minimum fetch length the waves are determined by the distance of open water the wind has blown over. In the area off the minimum fetch corresponding to the duration the wave dimensions are uniform, and the waves grow with time. The growth becomes gradually slower, and the waves asymptotically gain their optimal dimensions, which constitute the fully arisen sea, corresponding to the wind velocity. Both time and length required rapidly increase with wind velocity, the length most rapidly, and in most lakes a fully arisen sea is only developed by fairly weak winds.

In lakes the distance over which the waves are generated is generally approximately equal to the distance from shore to shore in the direction of the wind. For winds blowing along extended, limited waters this potential maximum fetch length is not equal to the effective fetch length. The waves at a certain point have been generated over an area and not along a line. The width of this generation area depends on the directional steadiness of the wind. In extended lakes the fetch area is thus dependent on the width of the water. In narrow waters the boundary conditions greatly influence the wave generation, and in a dammed, straight stretch of river, for example, the effective fetch is only a small fraction of the maximum length of open water.

For rectangular water surfaces relationships between the ratio of length and width and the effective fetch length, corresponding to areas of unlimited width, have been computed by Saville (1954). The problem of wave forecast for inland waters has been further investigated in a large research program directed by the Beach Erosion Board (1962). The methods and results of these investigations have been used in the calculations of the waves of Lake Vättern.

Wave characteristics

The wave height (H) is generally defined as the vertical distance between the wave crest and the preceding trough, but the mean of the difference of two consecutive crests and the intermediate trough is also used (Neumann 1953). The wave length (L) is defined as the horizontal distance between similar points of two successive waves.

In the stage of growing waves the maximum heights are more rapidly reached than the maximum lengths, and thus the wave steepness (H/L) is large and the sea gets its well-known "choppy" character. Stokes proved theoretically that the steepness can never exceed the ratio 1 : 7. Steeper waves become unstable, because the orbital velocity at the wave crest exceeds the wave velocity, which causes the wave to break. In nature, the waves generally break before reaching this steepness, but according to Bigelow and Edmondson (1947, p. 29) they may occasionally be as steep as 1 : 7.

The wave velocity (C) in deep water is primarily dependent on the wave length, but high waves of a given length run somewhat faster than low waves. From the definitions it follows that for all waves the relation

$$L = C\,T \tag{14}$$

holds true, where T is the period defined as the time difference between two successive waves.

In the theory of Gerstner and to a second order of approximation in the theory of Stokes, the wave velocity in deep water ($d > L/2$) is given by

$$C = (gL/2\pi)^{0.5} \tag{15}$$

and from the relation of (14), the wave velocity can be expressed as a function of T,

$$C = gT/2\pi = 1.56\ T\ \text{(m/sec)}. \tag{16}$$

In nature waves of a particular area or waves passing a fixed point during a particular time are not uniform but constitute a wave spectrum. The dimensions and the frequency ($1/T$) form distributions, from which statistical characteristics can be determined. The following characteristics and symbols are used as a standard (Wiegel 1953).

$H_{1/10}$, defined as the average height of the highest one-tenth of the waves (for a specified period of time).

$H_{1/3}$ or H_s, the average height of the highest one-third of the waves. It is also called the significant or characteristic wave height.

H_{av} or H_m, the average or mean wave height of all waves of the spectrum.

Analogous characteristics are used for the period (T). The significant height and period are the ones most commonly used to represent the wave spectrum in wave forecasting.

From wave records the ratios of the different characteristics of H have been found to be constant (Putz 1952), and almost identical relationships have been computed from theory (Longuet-Higgins 1952). These frequency distributions of wave height have been proved to hold for waves generated in limited water areas (Beach Erosion Board 1962, Pl. 3).

Water-particle motions

In the theory of waves of small (infinitesimal) amplitude and in the trochoidal wave theory the wave motion is purely oscillatory, and the water particles within a cycle (period) describe closed, circular orbits. The radii (r) of the orbits rapidly decrease with depth (z) below the water surface. In the trochoidal waves the radius is given by

$$r = 0.5H\, e^{-2\pi z/L}. \tag{17}$$

For waves of finite amplitude in the theory of Stokes the forward motion of the particles is not completely compensated by their backward motion, and thus the orbits are not closed. In addition to their oscillatory motion the particles have a progressive motion in the direction of wave propagation. The drift V_z at a distance (z) below the surface–experimentally verified by Mitchim (Beach Erosion Board 1942)–is

$$V_z = H^2 e^{-4\pi z/L}\ (g\pi^3/2L^3)^{0.5} \tag{18}$$

and the total volume transfer per unit width of wave crest

$$G = H^2(g\pi/32L)^{0.5}. \tag{19}$$

In wind-generated waves the drift near the surface is raised by traction of the wind force.

Waves in shallow water

The relationship between wave velocity and wave length for deep water in equation (15) is an abbreviation of the general expression for waves of infinitesimal height and trochoidal waves

$$C = (gL/2\pi \tanh 2\pi d/L)^{0.5}, \tag{20}$$

where d is the water depth and $\tanh 2\pi d/L$ approaches unity as $d \to \infty$ for a finite L.

If L is very large in comparison with d, $\tanh 2\pi d/L$ approaches $2\pi d/L$, and the wave velocity can be expressed by

$$C = (gd)^{0.5}, \tag{21}$$

where d is the vertical distance from the wave crest to the bottom. This is the expression for the velocity of solitary waves, also called translatory waves, because they involve a mass transport. The ideal solitary wave consists of an isolated crest raised above still water. There are no troughs, and the wave length is infinite. When the crest is passing, all particles are displaced forward the same distance, irrespective of the depth below the surface, and during the motion all particles at the same horizontal distance from the point of the maximum elevation have the same velocity, irrespective of the depth below the surface. There is no backward motion.

Waves in areas, where the depth is less than half the wave length, are generally called shallow-water waves in contrast to deep-water waves. This point of division has no specific significance as the influence of the depth increases continuously as the depth decreases. At depths for which $d/L = 0.5$, $\tanh 2\pi d/L = 0.9963$ and the wave velocity is thus less than 0.2 % lower than the velocity at an infinite depth.

In experimental studies (Beach Erosion Board 1948) no significant difference between the wave velocity according to equation (20) and the velocity of waves of finite height was found. As the theoretical formula of the latter is quite complicated, it has hardly been used in practice.

Shoaling waves

In geomorphology, as in coastal engineering, the behaviour of waves approaching shores and running up beaches is of central interest. The environment, in which the wave motion is directly influenced by the bottom, can be divided into three zones. The zone of shoaling waves, where the bottom has an appreciable effect, but no general breaking occurs, the breaker zone, where the waves regularly break, and the swash zone, where the waves run up the foreshore or beach face. Hence, the points of division between these zones on a particular shore are not fixed, but depend on the actual waves.

When waves enter areas of smaller depth ($d < 0.5L$), the wave velocity decreases, and because the period remains constant, the wave length is shortened proportionally to the change in wave velocity. If there is no or small loss of energy, a gain in height as a direct function of the decrease of wave length can be expected, but the wave form also changes. We find, especially from vertical air photographs, that the first visible effect of a shoaling bottom is a regularization of the wave crests in rows of coherent ridges, and these crest ridges appear to take up much less than half the wave length. The effect is a flattening of the troughs and a concentration of the water mass above the still water level.

According to the refraction theory (Munk and Traylor 1947), there is first a slight decrease in wave height in the interval $0.5 > d/L_0 > 0.15$, where L_0 is the wave length in deep water. Then the waves gain in height and for $d/L_0 < 0.06$ become higher than offshore. At its lowest the wave height is about 90 % of the height in deep water. The steepness of the waves continuously increases until they break. The breaker height of the waves, which are steep in deep water, is less than the deep-water wave height, because they break at a depth which is greater than $0.06L_0$.

The changes in the surface characteristics are due to the influence of the bottom on the water-particle motion, which is transformed from the bottom upwards. The orbits are changed into elliptical paths, which near the bottom are very flat.

It has been debated whether the wave motion off the breaking point should be treated according to the theory of solitary waves or that of oscillatory waves. The solitary-wave theory has proved to be of value in the prediction of breaking conditions (Sverdrup and Munk 1947), but its application to water-particle motion for waves which still have an oscillatory motion seems somewhat problematic. Some papers dealing with this problem will therefore be discussed.

The applicability of the second approximation of finite oscillatory-wave theory has been experimentally studied by Morrison and Crooke (1953). Most of the waves generated were rather steep. The maximum steepness was 0.096. Three different bottom slopes were used: 1:10, 1:50, and a horizontal bottom. The following statements are quoted from their conclusions.

The theory and measurement for surface time histories or profiles, for vertical and horizontal particle velocities, and for particle orbits show agreement within 20 % (based on measured data) for all waves in which $d > 0.2L$. In the interval $0.2L > d > 0.1L$, the theory showed fair agreement for waves of "appreciable steepness". For $d < 0.1L$ the theory did not agree with the measurements. In shallow water the horizontal particle velocity could be twice as high under the crests as under the troughs at the same elevation, but very steep waves over a sloping bottom resembled deep water waves in having the same horizontal particle velocity under crests and troughs. Low waves over a low beach were in this respect similar to very shallow waves.

In a paper by Inman and Nasu (1956), which deals with water-particle velocities, it is argued that the solitary-wave theory complements Stokes' theory. Their data are based on measurements of near-bottom velocities just seaward of the breaker zone at the Ocean Pier, Scripps Institute, La Jolla, California.

In order to obtain velocities consistent with the solitary-wave theory, which only involves motion in the direction of wave propagation, the maximum velocity under the preceding trough, or the mean of the velocity under the preceding and following troughs, were added (as a positive component) to the maximum velocity under the crest.

From their comparisons, the authors drew the conclusions that the observed velocities, as defined above, were in better agreement with the solitary-wave theory when:

1. The wave profile was not complex.
2. The effective wave period was longer than about 6 sec.
3. The wave height was higher than $0.4h$ to $0.2h$, where h is the water depth at the trough.

However, the diagrams of the relation between theoretical velocities and velocities, computed from the measurements, show a considerable scatter. In the diagrams, in which the velocity is correlated to the wave height, the theoretical relation can hardly be said to be visualized. Scattering could be expected from natural waves, but in the report it is also

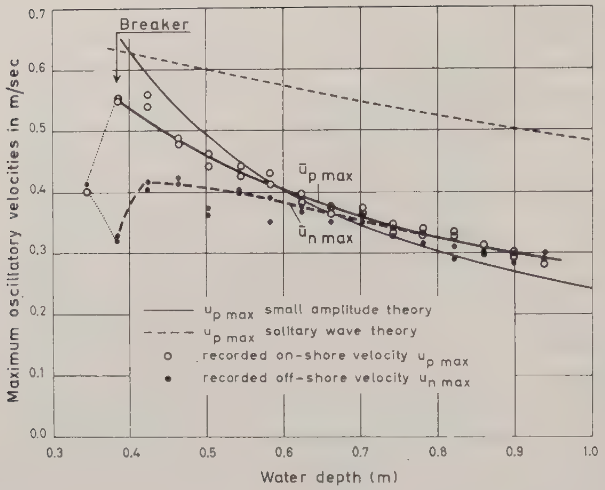


Fig. 36. Maximum onshore and offshore oscillatory velocities 5 cm above the bed for waves with $H_0 = 29.0$ cm, $L_0 = 5.0$ m ($H_0/L_0 = 0{,}058$). Bottom slope 1: 51. (From Johnsen 1961, Fig. 22.)

ascribed to the measuring technique, especially as regards water-particle velocity.

In a recently published dissertation on the relation between wave motion and nearshore bottom morphology, based on tank experiments (Johnsen 1961), the water-particle velocities off the breaking point are also examined.

The experiments were made on a large model scale. The dimensions of the waves ranged from wave lengths (L_0) of 2.70 to 7.08 m and heights (H_0) of 11.4 to 29.0 cm. The steepness in deep water varied between 0.031 to 0.058. The bottom had initially a slope of 1:51, and was covered with well-sorted sand ($Md = 0.16$ mm). The water-particle velocity was measured with micro velocity-meters of propeller type.

As regards wave velocity, the measured velocities agreed well with the velocities according to the equation for oscillatory waves (20), and were far from solitary-wave velocities. The change in wave height agreed well with the refraction formula of Munk and Traylor (1947).

The maximum onshore and offshore water-particle velocities, measured 5 cm above the bottom, were found to have the same magnitude and to increase at the same rate until a depth of $m\,H_0$, where m is a number which depends on the wave steepness and is larger for longer waves of the same height (op. cit., p. 26). From this point the onshore velocity increases more rapidly than the offshore velocity. The velocity under the crests increases to the breaking point and then falls off, while the velocity under the troughs reaches its maximum before the breaking point (Fig. 36). From Johnsen's diagrams it can be found that the point of velocity separation lies at $d = 2.6 H_0$ to 3.9 H_0 for $H_0/L_0 = 0.058$ to 0.031. In principle, the same relations between theoretical and measured velocities, as shown in Fig. 36, were obtained in all trials.

Johnsen does not compare his results with other investigations. If his maximum onshore and offshore velocities are totalled, one finds that the total velocity exceeds the theoretical solitary-wave velocities, and that the discrepancy becomes larger towards the breaking point. The theoretical particle velocity of oscillatory waves exceeds measured onshore velocities for $d < 0.11L$, except for the steepest waves, at which the limit lies at $d < 0.16L$ (these limits have been estimated from the intersections in Johnsen's diagrams, Abb. 21–25). The figures support Morrison and Crooke's statement of agreement with oscillatory theory for $d > 0.2L$ to 0.1 L.

From the data an empirical relationship between $C/u_{p\,max}$ and d/H has been determined, where $u_{p\,max}$ is the maximum horizontal onshore velocity 5 cm above the bottom (op cit., p. 30),

$$u_{p\,max} = 0.4C(H/d - 0.062)^{0.72}. \qquad (22)$$

With known deep-water characteristics, $u_{p\,max}$ can be determined for any depth by the computation of C from equation (20), and by taking $H = H_0$, or computing the height from

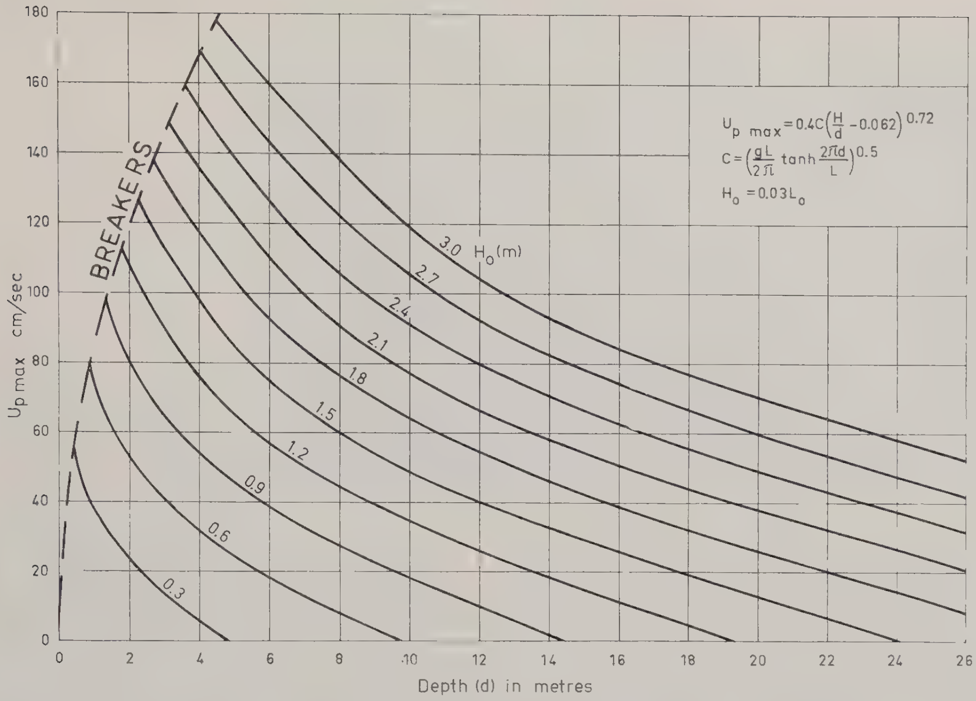


Fig. 37. Maximum onshore oscillatory velocities near the bottom for waves with a steepness in deep water of 0.03. According to formula (22).

the refraction formula. In Fig. 37 $u_{p\ max}$ has been calculated for waves with a steepness in deep water of 0.03.

It is not only the maximum velocities that are important but also the total velocity distribution. The difference in onshore and offshore velocity is associated with a difference in duration. With diminishing depth, the duration of onshore motion takes a decreasing part of the period, and hence the offshore motion an increasing part. This difference depends on the wave steepness, longer waves giving larger differences. The total effect of the velocity distribution in the vertical plane causes a mass transport, positively directed at the surface, negatively at the bottom.

Mechanically generated waves in tanks resemble ideal waves more than wind-generated waves in nature do, and the danger of generalizing results from tank experiments has been stressed (Inman and Nasu 1956). The scale in models is often so small that scale effects cannot be neglected. Because of the lack of wind stress, the waves are more symmetrical, and in closed tanks an onshore mass transport must be compensated by a reverse flow in the same section, which does not necessarily reproduce the natural conditions, where longshore movements are possible.

From the investigations quoted and from other papers it can be said, that, as far as short, steep waves are concerned, the motion off the breaker zone should be looked upon as following the theory of oscillatory waves, whereas long, low waves may be better treated according to solitary-wave theory, when they have become steep before the breaking point.

Breakers

In waves which are just about to break, the horizontal particle velocity at the crest is considerably higher than near the bottom, and also higher than the wave velocity. These relations can be ascribed to the decrease of wave velocity over a sloping bottom. If the bottom slopes very gently, the crests are still fairly symmetrical, but over a steep bottom the waves rapidly become assymetrical with steepened fronts. Breaking will occur when the depth is about 4/3 of the wave height.

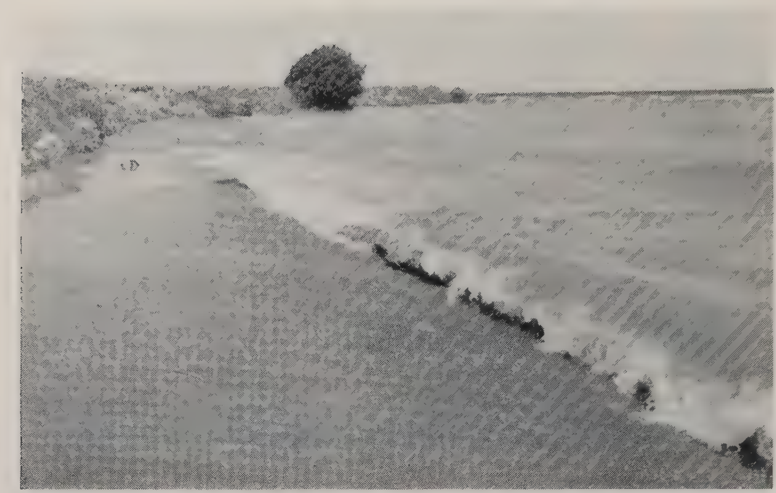


Fig. 38. A plunging breaker curls over on the step. The shore west of Sandudden. July 1957.

There are two main types of breakers, which have been called plunging and spilling breakers. In the former type the water curls over the vertical front and falls down. This type, which is generally formed over a steep, even bottom or at a step, implies a large momentary loss of forward-directed, kinetic energy. After the breaking, a wave of less height may be reformed, again breaking at a smaller depth. A spilling breaker, on the other hand, advances at the correct speed for the depth with a foaming crest; the wave does not lose its identity, but gradually decreases in height until it becomes swash on the beach (King 1959, p. 15). This type of breaker resembles solitary waves. They are generated on gently sloping bottoms, especially from a long, low swell, which steepens and forms rows of breaking, coherent ridges off low, sandy beaches.

The breaking transforms kinetic, directed energy into a turbulence of confused eddies. Over a bottom of loose material, a part of the turbulent motion expends itself in erosion and transportation of solid particles, but turbulence also means an energy transformation to heat.

The eddies move at a speed considerably lower than their internal rotational velocity. The tracks of the eddies are influenced by the currents of the breaker zone. The internal eddy velocity is important, because it involves a lifting force which is much larger than that of the average current velocity.

The general relationships of the water-particle velocities within the breaker zone in

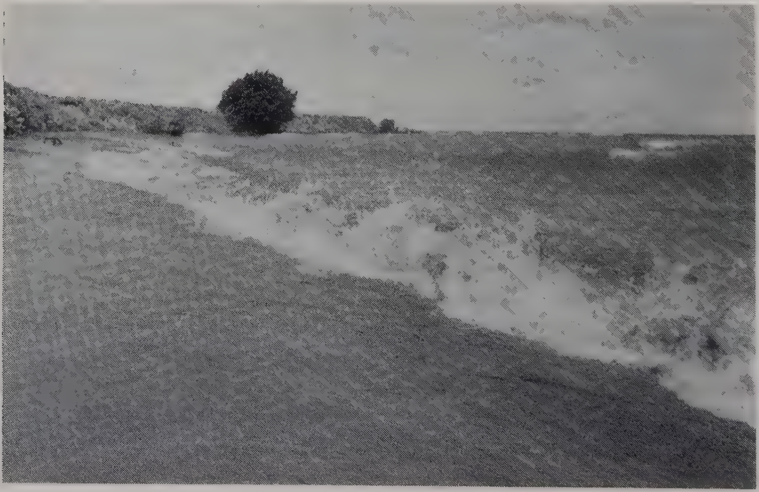


Fig. 39. The front of a plunging breaker falls down. Note that a thin layer of back wash still covers the foreshore.

relation to particular wave characteristics in deep water and particular bottom conditions have as yet not been established. Because of the large velocities and forces it is very difficult to work in this zone, and no adequate equipment for measuring and recording the turbulent motion has been available. In a large field and laboratory research program on shore processes, which was planned at the Coastal Studies Institute, Louisiana State University, and started in 1963, the breaker zone is one of the main objects. In this project direct measurements in the breaker zone under natural conditions have been planned.

With the use of Scuba equipment it is possible to get a fairly good qualitative idea of the general motion in the breaker zone. In front of a wave ridge the offshore motion rapidly accelerates to form the crest. Just as the crest passes, a short onshore motion comes as a blow, but as soon as it has passed, the water is filled up with eddies, and it is generally difficult to distinguish any specific onshore or offshore motion until in front of the next wave, when the motion is again regularized to a real surge. This description, which is based only on direct observation, is identical with some of Inman and Nasu's records of the motion just before breaking (1956, Fig. 14).

Swash

The swash starts where the wave runs up the beach in a true mass movement. This uprush can be of the plunging (Fig. 38) or spilling breaker type. The uprush is very turbulent, and the onshore-directed energy is transformed into turbulence, the work of transportation of solid particles, and the potential energy of percolated water and surface water.

The downwash starts when the onshore-directed, kinetic energy has been expended. Some eddies may be left, but the essential feature is that, in contrast to the uprush, there is very little turbulence when the downwash starts. The turbulence of the wash is formed during the outward motion. The downwash ends in a situation in which either all the water in front of the next breaker is gone or there is a sheet of water left.

Like the breakers, the plunging swash is found on steep beaches, and the spilling swash on low beaches. The primary importance of the slope has been demonstrated by records of the water-particle motion of waves, breaking on an even bottom of different steepness (Iverson 1952).

Sometimes a third type, the surging swash, is distinguished (cf. Beach Erosion Board 1961, Fig. A-4). The uprush of this type sends up a low tongue of foaming water. It is associated with very steep beaches and waves of rather low steepness. The most extreme surging swash is found on solid structures and smooth, rocky shores, such as glacially eroded sheep backs.

In inland waters with short, steep waves and high beaches, the plunging swash predominates. The plunging at the start of the uprush means a momentary breakdown of the wave form and an eddy formation in a narrow zone. During the uprush the water advances with a rather horizontal surface, which means that a large part of the water mass stays in the lower portion of the swash zone. In the downwash the water of this portion is rapidly lowered, and then flows down in a thinning sheet (Fig. 39).

On low ocean beaches, especially in hard weather, the up-and-down motion is not so well distinguished; the breakers are broken down to spilling water in a broad zone, the uprush is spread over a large area and is not withdrawn before the next foaming crest comes riding in.

Refraction phenomena

The decrease of wave velocity and wave length with diminishing depth has previously been mentioned. If we look upon a system of coherent wave crests, which obliquely approaches a sloping bottom with straight contours, it is obvious that the part of the system which is nearest to the shore will have slowed down more than the parts which are in deeper water. The crests thus tend to swing into a direction parallel to the bottom contours. This refraction follows the general refraction formula of all kinds of waves,

$$C_1/C_2 = \sin \alpha_1/\sin \alpha_2 = L_1/L_2, \qquad (23)$$

where C_1 and L_1 are the incident velocity and wave length, C_2 and L_2 the refracted velocity and wave length, α_1 the angle of incidence, and α_2 the angle of refraction.

Methods of practical refraction analysis have been summarized by the Beach Erosion Board (1961). In these methods it is assumed that no energy flows laterally along the wave crests. The direction of wave propagation is represented by orthogonals, which are lines drawn perpendicular to a wave crest as it passes over the changing hydrography.

If the bottom contours are straight lines and the wave crests, before they break on the beach, turn into a direction perpendicular to the contours, the concentration of wave energy per unit width of wave crest will be reduced by cos α_1, which could be expected from pure geometry. Where the bottom contours are convex towards the water, the orthogonals will converge. Orthogonals, which in deep water have been drawn at a particular distance from each other, will thus come closer, and with the assumption of no flow across the orthogonals, this means a concentration of energy. Over a bottom with straight or concave contours with diverging orthogonals there will consequently be a deconcentration of energy.

The change in concentration of wave energy per unit width of wave crest means a change of wave height. Munk (1949) has introduced a factor, called the refraction factor K, by which the wave height is increased or decreased by refraction. Waves, which are not about to break, are treated according to oscillatory wave theory, in which the wave energy is proportional to the square of the wave height, and K can thus be expressed by

$$K = (S_0/S)^{0.5}, \qquad (24)$$

where S_0 is the initial distance between the orthogonals and S is the distance between the same orthogonals of the refracted wave.

Waves which are about to break are treated according to the solitary-wave theory, for which the energy is proportional to the cube of the wave height, and thus the refraction factor (K_b) of such waves is

$$K_b = (S_0/S)^{1/3}. \qquad (25)$$

Although the assumption of no flow across the orthogonals is not strictly valid, good agreement between wave heights predicted from theory and observed wave heights has been found to exist. From the refraction it follows that, in the case of a sloping bottom, more energy is transmitted to projecting parts of the shore than to bays, and that the rise in water level will be higher in the former areas.

Currents induced by waves in shallow water

Mass transport by wave motion and return flow

The wave motion involves a certain transportation of water mass (equation 19), and by frictional drag of the wind there is also some additional surface flow in the direction of wave propagation. This means that on the leeward shore, where the transportation is blocked, a water-level gradient sloping off the shore will develop. Because of wave refraction, the gradient is not the same all along a shore. The water is raised higher in areas of wave convergence than in areas of divergence, and thus there may be a gradient along the shore. Waves obliquely approaching a shore cause a mass transport with an alongshore component.

Where there is a water gradient, there is always a flow in the direction of its slope. Earlier the model circulation of large drift currents, with an onshore drift at the surface and a reverse drift along the bottom, was believed to form the general pattern of the leeward nearshore area affected by wave-induced currents. As pointed by Shepard (1948, p. 43), this undertow may exist, but its strength has often been overestimated.

In 1940 systematic measurements of mass transport were carried out in a wave tank by Mason (Beach Erosion Board 1948). At the same time the theories of oscillatory waves, including mass transport in shallow water, were summarized by O'Brien and Mason (1942). According to the theory of Stokes, the rate of drift in shallow water for a certain wave height rapidly increases near the surface with decreasing wave period, whilst the rate near the bottom only slightly decreases (op. cit., 1942, Fig. 7).

In a closed tank the onshore mass transport must be compensated by a reverse flow in the same section, which reduces the net flow to zero. The theoretical net motion was obtained by assuming the return flow to be uniformly distributed from the bottom to the surface. In this, from a dynamic point of view questionable assumption, the form of the onshore velocity profile was not affected by the reverse flow. It only meant that the origo of the system was horizontally displaced. The theory of Stokes does not take boundary conditions into consideration, nor does this model of net motion.

Mason's experiments proved the existence of mass transport. At depths greater than $0.5L$ the empirical results agreed quite well with the theoretically predicted net-velocity profile (velocities close to the bottom seem not to have been measured), but in shallow water ($d<0.5L$) the results were ambiguous (op. cit., Fig. 14).

In tank experiments, using dye as a tracer, Bagnold (1947) found that there was a slow seaward drift in the upper parts of the tank, but a strong shoreward movement on the bottom, as long as the bottom was smooth. The presence of ripples was associated with turbulence, which seemed to prevent the forward movement of the dyed bottom water. When the dyed tongue of water approached the break-point, the tongue moved vertically upwards and became dispersed by the turbulence. Near the surface the dyed water started to move off the shore. Dye dropped inside the break-point also became dispersed, but did not move seaward of the break-point. Almost the same results from model experiments in Cambridge are reported by King (1959, p. 102). In these experiments the motion off the shore seems to have a maximum at about mid-water depth. King adds the supplementary information that in deep water there was still a landward movement on the surface. She also states that the experimental data agree with the theoretical model by Longuet-Higgins (1953), in which the conditions of the boundary layer are taken into account. About the onshore movement on the surface it can be remarked that this can be expected to be larger in wind-generated waves than in the paddle-generated waves of a tank. It should also be remembered that in nature lateral movements are not restricted as in a narrow tank, and because of refraction onshore mass transport is not uniform along the shore.

One of Bagnold's previously mentioned observations, which is of great importance, is the discontinuity of the movement at the break-point. It can be interpreted as giving a damming effect within the breaker zone.

From theory and experiments different modes of reverse flow seem to be possible in nature: continuous, laterally rather uniformly distributed drift in the middle and upper layers of the water mass, longshore drift in the direction of wave approach and from areas of high energy towards areas of low energy, and concentrated outflows from the breaker zone in areas, where the damming exceeds a critical value.

Longshore currents

A formula for the estimation of longshore-current velocities (V) inside the break-point has been devised by Putnam, Munk and Traylor (1949):

$$V = K\left[(mH_b^2/T)\sin 2\alpha_b\right]^{1/3}, \qquad (26)$$

where $m = d_b/l$ (d_b is the breaking depth, l the distance from the break-point to the shoreline), $K^3 = cg(s_t/k)$ (c is a constant, s_t is the sand-surface texture, and k is a function of the hydraulic roughness), and α_b is the angle between the wave crest at the break-point and the shore.

The principles of the formula are quite simple. The wave velocity and energy per surface unit are obtained from solitary wave theory, and are represented by H_b and T. The distance over which energy is lost by friction is represented by m and d_b. The longshore component outside the break-point is a function of sin α_b, and the fraction of this per unit width of the breaking-wave crest is a function of cos α_b. Thus, soi $2\alpha_b$ is the composite result obtained from the relation $\sin 2\alpha = 2 \sin \alpha \cos \alpha$.

From $\sin 2\alpha_b = 1$ for $2\alpha_b = 90°$ it follows that the current velocity has a maximum for $\alpha_b = 45°$, and that the velocity distribution is symmetrically distributed around this angle and is zero for $\alpha_b = 0°$ and $90°$.

The quantitative relationship has been found by empirical determination of K. Rather large variations can be expected in nature. The waves are assumed to be solitary waves. The formula only applies to a straight beach of uniform bottom slope, and the texture should also be uniform. The formula is not included in the standard formulas for beach-protection planning issued by the Beach Erosion Board (1961). Still the formula clearly demonstrates how the current increases with increasing breaker height and bottom slope and decreasing wave period.

Formulas like equation (26) have been used in forms applied to a direct estimation of the rate of longshore drift of sand (cf. Savage 1959).

$$E_a = E_0\, 0.5 \, \sin 2\alpha_b \qquad (27)$$

and $M = E_a^k$, (28)

where E_a is the alongshore energy per time-unit per length of beach, E_0 the wave energy per time-unit, M the volume of sand transported and k a constant. That k is a constant is, of course, not really true. It must be a function of bottom morphology, sand texture, wave types, etc. Saville (1950) has shown that the littoral transport rate is a maximum for a deep-water wave steepness of the order of 0.025, and that for steepnesses larger and smaller than 0.025 the transport rate decreases.

The discussion on the relation between wave incidence and maximum longshore drift is somewhat confused, because alternately a_0 (the angle of incidence in deep water) and a_b is used.

Table 11. Wave incidence of maximum longshore drift.

α_0	α_b	Reference
	45°	Formula (26) and (28)
43°		Shay and Johnson (1951),
30°		Johnson (1952)
53°		Sauvage and Vincent (1954)
54°	(24°)	Bruun (1954)

About the 53° angle, found in a laboratory study by Sauvage and Vincent, Savage (1959, p. 4) remarks: "It is not clear whether this is the breaking angle, as is implied in the text, or the deep-water angle of wave approach, which appears more logical from the size of the angle". The angle of 30°, found by Johnson, is not commented on, although it cannot be co-ordinated with the other results.

The angle in deep water, which results in the strongest longshore drift, is the one which maximizes the summed energy of the refraction before breaking and the longshore component of the breaking wave.

According to Bruun (1953, p. 13), the refraction before breaking is taken into account in a formula, in which the drift is proportional to $(\sin a_0 + \sin 2a_0)$, which gives a maximum for $a_0 = 54°$. The formula is based on a deep-water wave steepness of 0.04. With this steepness the angle at the break-point is about 24° (op. cit., Fig. 26). The formula is said to be valid for waves of a steepness > 0.01. One finds that, when the waves in deep water approach the maximum possible steepness, a_b approaches 45°.

An important consequence of Bruun's formula is that, with a wave propagation parallel to the shore in deep water, the littoral drift is still 60 % of the maximum drift. The energy is not totally dispersed by this refraction in nature, but it seems uncertain, whether as much is left as Bruun's formula indicates.

The intensity of alongshore transportation by waves of a low angle of incidence is of importance in Lake Vättern, where the strong winds blow along the lake. It shows the small possibility of permanent deposition of sand on the long, rather straight, easterly and westerly shores.

Nearshore circulation patterns

The existence of longshore currents does not explain how the onshore mass transport is compensated. From bathing beaches it is known that swimmers can be caught by strong seaward currents. These currents have been mistaken for an undertow, but they really flow at the surface. The most extreme type forms a narrow lane through the breakers, and in the deep water it dissipates in a fan of macro-vortices. This type is called a rip current. For the return flow in general, which is restricted to a narrow lane, Davidsson (1963, p. 33) has proposed the term "cross-current". Rip currents have been described in several papers, among which Shepard, Emery, La Foud 1941, Shepard 1948, Shepard and Inman (1950), and McKenzie (1958) should be mentioned. These currents are fed by along-shore circulation in the nearshore zone, and their strength depends on the dammed water mass, which varies with changes in wave height. Along rather straight beaches, the position of the rip currents varies from time to time, but they are usually more permanent near obstacles such as a jetty or a natural projection of the shoreline.

Extensive studies of nearshore currents by the use of floats have been carried out on Scripps Beach in California (Shepard and Inman 1950). In these studies various circulation patterns within the nearshore zone were found to exist. Systems of large circulation cells with narrow outflows between them seem to be most typical.

A nearshore circulation pattern, recorded on the north-western part of Visingsö, is illustrated in Fig. 40. The currents were traced by injections of fluorescine and by observation of the

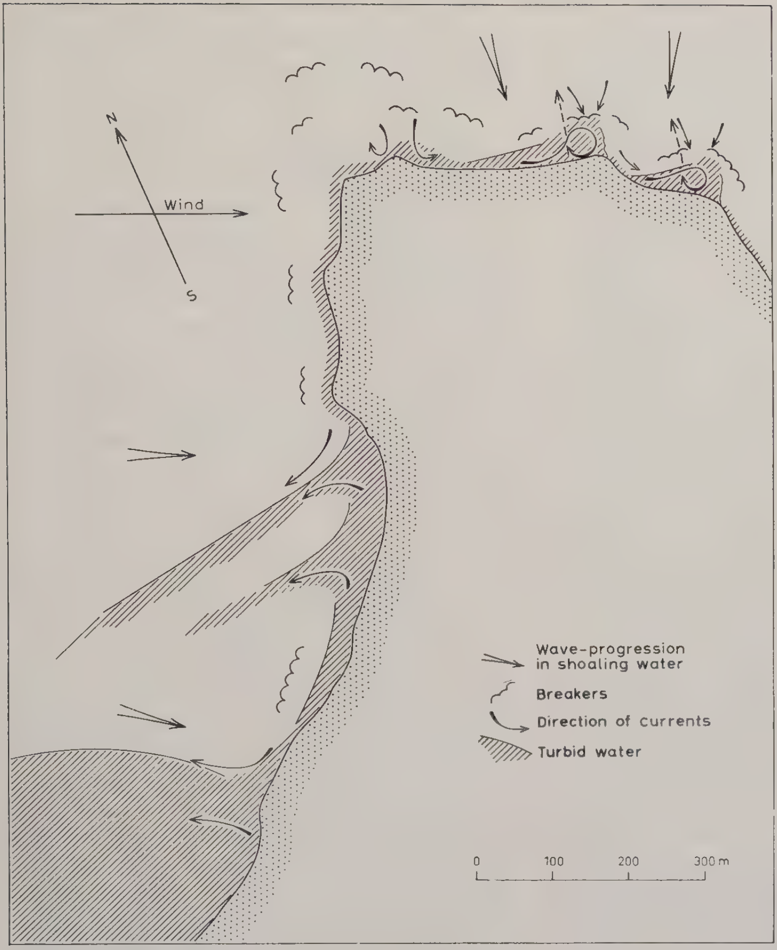


Fig. 40. Nearshore circulation at the northern end of Visingsö island on August 2nd, 1957. Wind velocity 6–8 m/sec.

flow of turbid water dispersed in the breaker zone. The wind was blowing from the west. The wind velocity measured at 4 m above the water surface varied between 5.6 and 8.4 m/sec, and the significant wave height was estimated to 0.4 m.

On the northern shore the wave refraction was 90° and even more on the heads between the bights. Along the bights the current was easterly and followed the shore, but at the down drift end the current turned off the shore. On the heads the current was directed towards the shore, and thus a circular current with a diameter of 30–40 m developed. The dyed water passed the circles in about 35 min, which means an average velocity of about 5 cm/sec.

Suddenly, there was a break-through and a flow of water across the breakers, as indicated by the dashed arrows. Then the closed circular movement continued again with only slight leakage through the breakers. The point of outflow is situated, where the longshore current meets the current from the high energy zone of convergent breakers at the head.

On the western shore the outflows were also concentrated to the flanks of projections. Here, no circular movements were found.

Some observations of the flow of turbid water within and off the breakers have been made by Scuba dives. In the breakers the suspended material is distributed from the bottom to the surface, but in the zone off the breakers almost clear water has been observed near the bottom. This is in agreement with the previously mentioned laboratory experiments by Bagnold and King. Even though there is almost zero visibility in the turbid water, the density of the silt and clay suspension is not large enough to form a density stratification dependent on the suspended material. The observed distribution of the suspended material can to some extent be explained by a difference in temperature between outflowing water, warmed at the surface, and inflowing colder water along the bottom. When there is suspended material in the surface water off the breakers, this can be explained by the turbulence of waves breaking in deep water.

The waves of Lake Vättern are generally steep and short, and there is no regular broad surf zone with coherent rows of breakers. Often the only common break-point lies at the outer limit of the swash zone. From several experiments with dyed water the author has found that dye, injected in the swash, disappears very slowly from this zone, especially when there is a pronounced longshore drift (cf. Russell and Macmillan 1952, p. 121).

Davidsson (1963, p. 33) states: "Between the coastal current and the nearshore circulation there is no sharp limit. When depth increases particularly rapidly the coastal current can extend to a great part of the breaker zone." The coastal current is the slow movement in the offshore area, caused by wave mass transport, wind and unequal air pressure. In my opinion, the existence of a sharp limit depends on a regular break-point, because experiments and observations in nature have shown that a double circulation system exists, separated by the break-point (King 1959, p. 103). When the regular breaker line lies at the swash limit, the inner zone is certainly narrow, but there may still be a sharp limit.

Russell (Russell and Macmillan 1952, p. 104) points out that, although the strong off-shore currents of the nearshore area only appear in the form of rip currents near the surface, still a slower undertow probably often exists. It may be so, but the proofs to which he refers are not satisfactory. In experiments with bowling woods, weighted so that they move along the sea bed, and in other experiments with water-filled balls, the objects were not recovered on the beach. The conclusion is that they were carried off the shore by a current of the same direction. The difficulty is that those objects have not the same inertia as the water, because they form a rigid volume. Where the total mass transfer towards and off the shore is in balance in shallow water, that is to say, the net motion is zero, the objects may still move from the shore. The water-particle velocity under the wave crests is high, but the duration is very short; under the troughs the relation is the reverse. Because of their inertia the objects only gain a fraction of the particle velocity under the crests, while in the longer and more slowly accelerating movement under the troughs they gain a proportionally higher velocity. The net direction of motion of the water is thus not necessarily the same as that of the objects on the bottom. The relation depends on their inertia and the velocity distribution of the water particles.

Morphological activity

The critical erosion velocity

The importance of the concepts 'erosion', 'transportation' and 'deposition' in relation to the current velocity of a stream has been clearly proved by Hjulström (1935, 1939). The velocity necessary to initiate a motion—the critical erosion velocity—increases with the grain size for particles larger than about 0.2 mm. The velocity necessary to initiate the motion of a solid particle at rest on the bottom is higher than the velocity necessary for its further transportation. The relation between the critical erosion velocity for a flow parallel to the bed and the grain size of a well-rounded material of uniform texture (grain size > 0.2 mm) on a horizontal and sloping bottom has been theoretically determined and also empirically verified for a horizontal bottom (Sundborg 1956).

The critical erosion velocity for flake-shaped material and non-uniform material is in general still unknown. It is known that in a distribution, which has a large fraction of a grain size with a certain critical erosion velocity and a smaller fraction of a grain size with a somewhat higher necessary velocity, the latter will be eroded and transported at the current velocity corresponding to the finer grain size, and the coarse particles can move faster than the fine ones. However, if the coarse particles are much larger, they will only move occasionally, if the current velocity is far below their specific critical erosion velocity, and during a stage of net erosion they may form an erosion pavement.

For a sloping bottom the frictional resistance in relation to a flow parallel to the bottom and perpendicular to the bottom contours is theoretically proportional to tan Ø cos $i \pm \sin i$, where Ø is the angle of repose, the bottom is inclined at angle i, and the minus sign refers to a flow down the slope (Sundborg 1956, p. 174).

As regards the resistance to erosion of very fine soils, the frictional and buoyancy forces are negligibly small, while the cohesive forces are important. The fine particles stick together by mass attraction, intermolecular forces and electrostatic forces (cf. Grim 1953, p. 9). It is well known that, if the orientation of flaky clay particles is disturbed, the bindings will be looser, and also that a high compaction reinforces the bindings. The sediment is thixotropic. Loosely packed, fine sediments have a very different resistance to erosion than sediments with a low porosity. Thus the resistance cannot, as in frictional soils, be directly related to the grain size. Even if a numerical factor of the cohesion is introduced (cf. Terzaghi and Peck 1948, p. 82), this is of small practical value, as long as no field method for the determination of cohesiveness in relation to resistance to erosion has been developed.

In clayey sediments the bindings between aggregates of various sizes are generally much weaker than the bindings between the single particles. There are structural joints, and in stratified sediments there are weak zones at the bedding planes, especially where there is some coarse material. Sedimentary Glacial and Post-Glacial "clay" is therefore generally eroded in flakes along less resistant strata or along fracture zones (Arnborg 1958, p. 11, Norrman 1963*b*, p. 33). Where frictional material invades fine sediment surfaces, the corrasive (grinding) action of the sand and gravel starts at far lower current velocities than the flake erosion, and the corrasion is much more effective. This has been found in field studies and in preliminary laboratory studies at the Department of Physical Geography in Uppsala. The corrasion process is further discussed in the chapter on abrasion.

Modes of transportation

The definitions and types of movements of material under fluvial conditions have been summarized by Sundborg (1956, pp. 204–205). He states that the classification should be: contact load, saltation load and suspended load, or bed load and non-bed load. The latter term covers all suspended material that is lifted above the lowest layer of the stream. The contact load includes sliding and rolling particles. Sundborg points out that the saltation–the jumping movement of particles, articulated by elastic bouncing–is much more important in air than in water, because in air the velocities are higher, and the loss of weight due to buoyancy is negligible.

The suspended load consists of particles kept in suspension by the turbulent motion. By vertical turbulent exchange particles are brought upwards against the gravitational force, but particles are also brought downwards. There is always an exchange of particles, even if the concentration at each level is fairly constant. It has been possible to calculate the percentage of material carried in suspension and bed load in relation to grain size and current velocity (Sundborg 1956, Fig. 23, 1963, Fig. 34). These calculations refer to uniform material and steady tranquil flow.

It is of great morphological importance, both in fluvial and in shore processes, that there is a limit to the size of the finest material, which can be transported as bed load. Inman (1949, p. 57) found this size to be 0.18 mm, and in Sundborg's diagrams it lies in the interval 0.15–0.2 mm. Sundborg (1956, p. 219) stresses that the limit is diffuse and depends on the flow conditions.

Because of the fact, already demonstrated by Hjulström (1935), that the critical erosion velocity is far higher than the velocity associated with the cessation of the movement of fine particles, it follows that at velocities lower than the critical erosion velocity, corresponding to the grain size of about 0.15 mm, neither erosion nor transport of bed load can occur, but transport and deposition of fine suspended material may occur.

The flow conditions near the bottom under waves differ from the fluvial conditions in the oscillatory forward and backward motion, the large and rapid velocity variation and the great macro-turbulence (especially in breakers and swash) in relation to the mean velocity. Other conditions which make the shore environment more complicated are the relation between true wave motion and wave-induced currents, the two-dimensional movement over a sloping surface and the abrupt changes at break-point and swash.

The high momentary velocity under the crests of waves which are about to break and under breakers is associated with the development of vortices, which can lift particles far above the bottom. Empirical investigations (Watts 1953) have shown that the maximum concentration of suspended material (the swash not included) is found just inside the breakpoint. There is a rapid deceleration of the water motion behind the crest, but, if particles have been lifted from the bottom, even the slightest current will cause a horizontal displacement during their gravitational fall. By pure wave motion this will cause a backward and forward motion, perpendicular to the wave crests, but any slow current with a diverging direction will introduce a sideward component of motion.

All investigations on the amount of suspended material in the nearshore area agree in a rapid decrease from the bottom and upwards and in an increase of grain size towards the shore (cf. review by Emery 1960, pp. 240–241). However, no general relationship has yet been established between the characteristics of waves, grain size and other bottom conditions. The great difficulty is that there are no accurate methods of sampling in the surf (cf. Watts 1953) nor of measuring the turbulent motion.

A narrow range of grain sizes in the zone off the swash is a characteristic feature on most large sand beaches. There is obviously a predominance of medium and fine sand. Trask (1955), for example, reports a median diameter of 0.165 mm (2.6 Ø) on samples taken between depths of 0 and 40 ft along the Californian coast, and Davidsson (1963) gives the median 0.27 mm (1.8 Ø) of samples from the breaker zone on the Scanian coast in the Baltic. These grain sizes lie within and close to the interval in the limit of bed-load transportation.

A diagram by Sundborg (1956, Fig. 34), in which the relations between grain size, critical erosion velocity, velocity corresponding to the cessation of movement, and relative concentration of suspended load are given, explicitly demonstrates the great dependence of grain size on the mode of transportation and the rapid transitions in the sand fraction. In the interval 0.15–0.4 mm the critical erosion velocity is about 0.35–0.40 m/sec (the velocity measured 1 m above the bed), and the velocity corresponding to a cessation of movement is 0.25–0.27 m/sec. Thus these velocity ranges are small, but the corresponding range of the maximum velocity of a true bed load is 0.25–0.80 m/sec. A great relative concentration of suspended load requires far higher velocities; a concentration of 0.5 requires a velocity of about 3.0 m/sec for 0.2 mm particles. These figures from the diagram clearly tell us that sand of about 0.2 mm in size is generally too fine to roll along the bottom and too coarse to be kept steadily in suspension.

Under the waves this material will be intermittently lifted and deposited. The particles can be said to make jumps. However, this is not saltation but an intermittent suspended load. It is typical of all exposed shores that there are no fine sediments, and the reason is clear. All material which does not settle in the intervals between the maxima of the oscillatory motion will move out of the environment, that is to say, into deep water.

The depth of direct wave action is commonly determined by the occurrence of oscillatory (symmetrical) ripple marks. The details of the formation of ripples have been studied both experimentally (Bagnold 1946, Scott 1954, Manohar 1955) and in the field (Inman 1957, Zenkovitch 1962). (Model studies, in which the bottom has been oscillated in relation to a water mass at rest, are somewhat questionable from a dynamic point of view.) Two types of movements associated with ripple formation have been distinguished; a rolling movement, by which grains are arranged in very low rows, and at higher velocities an interaction between vortices in the water (lifting material) and much steeper ripples (Russell 1952, p. 114). Manohar (1955) states that ripples only form when the flow of the boundary layer is turbulent. According to the previous discussion on bed load and suspended load, ripples are not likely to form in fine material. An oscillation of a very loose, fine sediment, which behaves like a dense fluid, can probably produce ripple-like forms, as rows of small ridges can be artificially produced by vibration, but it is doubtful if these should be called ripples.

Wave action, which directly affects the bottom, can be expected to produce a separation of material belonging to the bed load, the intermittently suspended load and the permanently suspended load. Because of changing velocities the grain sizes of different modes of transportation will change, except for the lower limit of rolling movement. In an area, where no new material is introduced, particles which can be kept in suspension can be expected to be entirely separated from the zone of direct wave action.

In areas where fine sediments are abraded, or where fine particles are brought to the shore from rivers, there will be a continuous separation. At a certain depth the oscillatory current velocities change with time. During calm conditions after a period of hard weather the nearshore water is turbid, and the fine particles settle in areas where a short time before this had been impossible. The distribution of this deposition has no specific relation to the depth or the distance from the shore. The finest suspended material can be carried far even by a very slow current. Emery (1960, p. 241) reviews reports of divers, who had observed "a kind of haze" near the bottom in shelf areas, and I myself have observed how a layer of turbid water can for a long time form a density stratification in depressions in the bottom.

If the suspended material is deposited at depths, where there will be erosion during the next period of bad weather, it will again form a suspension. However, the fine material may prohibit erosion. Suspended material which settles on a sand surface fills out the pore space between the sand grains, and even a very thin layer is enough to give the surface a cohesiveness and a lower degree of roughness.

The same condition can be found in rivers. The regular formation of a "winter coat" during the low stage of the Ångermanälven river has been reported by Arnborg (1959*a*, pp. 103, 171). These thin layers of fine sediments often hold an appreciable amount of organic material. The cohesiveness raises the critical erosion velocity above that of the embedded sediments.

The best example of this process I have seen on any shore is in the delta of the Indalsälven river. The morphology is dominated by wave action. The general material consists of coarse sand with some granules, but during periods of calm weather and probably mostly during periods of ice ocver, fine sediments settle in the nearshore area. Even if the layer is less than 1 mm, and thus does not totally cover the coarse material, the surface is not eroded until pieces a few square centimetres in area are broken off or rolled up. The surface can be invaded by the coarsest available material without being destroyed.

On shores, the process described is most important in the zone of the largest possible depth of direct wave action, where the probability of transportation and deposition is much larger than the probability of erosion. The problem is further discussed in the chapter on the sediments of Lake Vättern.

The null-point hypothesis

Introduction

On a shoaling bottom, the net movement of material influenced by waves depends on the balance between the shoreward and seaward velocity distributions of the water particles and the angle of the slope. If the velocity distributions are identical, a net seaward movement of all solid particles in motion is expected because of the slope. With balanced but increasing velocities towards the shore, the coarsest particles would stop in the shallowest water, and the result would be a zonal sorting. If the effect of slope could be neglected, there would be no net movements at all but an oscillation of the particles.

Natural wind-generated waves are asymmetrical, and the asymmetry of the water-particle velocity distributions is pronounced in shoaling water. The maximum velocity is higher under the crest than under the trough, but the duration is longer under the latter. This means that for a certain wave and depth, the shoreward flow can be expected to initiate the motion of a particular maximum grain size but not a reverse motion of the same grain size, if the lower maximum velocity under the trough is not compensated by less resistance to erosion because of the slope. The deeper the water, the smaller the particles that can be moved by a certain wave size and, because the difference between shoreward and seaward velocities decreases with depth, the less chance there is of shoreward transport.

From this model it could be expected that a particle which has started a shoreward net motion will continue in the same direction. According to a hypothesis due to Cornaglia 1887, it will do so, but he also stated that, seaward of the point where the shoreward net movement starts, particles of the same size move in a seaward direction. The point of zero net motion–or rather the point of diversion–is called the null-point or the "neutral line". Each particle size has a specific null-point for a certain bottom slope and wave characteristic. Shoreward of it the particle moves shorewards, and seaward of it seawards. However, if the slope is low and the particles large, they cannot be moved, except in very shallow water. There the difference between onshore and offshore velocities is large, and it seems improbable that there will be any offshore motion. On the other hand a steep slope, small particles and growing depth increase the conditions favourable to an offshore movement.

Experiments

No real improvements in the null-point hypothesis were made until recent years, when it was theoretically examined by modern hydrodynamic theory and also experimentally investigated in a flume at the Massachusetts Institute of Technology (Ippen and Eagleson 1955, Eagleson, Peralta and Dean 1958). There are two reports of special interest from the Massachusetts studies. One relates the grain size at different depths to the wave size for a bottom slope of 1:15.

$$w = 11.6^{-1} HLCd^{-2}. \qquad (29)$$

The grain size is represented by w, which is the terminal fall velocity of the sediment particle in fresh water.

The other formula gives the relation between the median sediment diameter and the slope on equilibrium beaches.

$$D^{7/4} \sin i = \bar{U}_t T^{-0.36}, \qquad (30)$$

where D is the diameter, i is the angle of inclination of the slope, and $\bar{U}_t$ is Stokes' fluid mass-transport velocity. The dimensions of the formula are given according to the metric system. The two formulas refer to the area seaward of the breakers, and they hold for wave steepnesses of $0.01 < H_0/L_0 < 0.08$.

Formula (29) will first be discussed. It may be noted that the constant (11.6) is only valid for a slope of 1:15, and this is steeper than most slopes found in nature off the surf zone. The experiments were made with rather coarse material and few particles moving on a permanent flat surface roughened by adherent sand of a size less than the loose particles. Ippen and Eagleson's conclusions (op. cit., p. 79) may be quoted:

(1) The null-point was found to exist.

(2) Sediment particles are placed in oscillatory motion, whenever the maximum instantaneous fluid velocities near the bottom exceed the velocity necessary for incipient motion of the sediment particles.

(3) No sediment particles can move through the breaker into the surf zone under the type of motion studied.

(4) A particle in position offshore of its null-point under a given wave will move offshore, until it enters a region where the fluid velocities are reduced to values of the incipient velocity. This limit has not been fixed as yet by the experiment.

Applications to natural shores

As far as is known, curves of the null-point grain sizes corresponding to different waves according to formula (29) have only been published by Miller (1958). He presents five curves of waves of a length less than 15 ft, and still the particles are amazingly coarse. In order to illustrate the relation for waves of natural sizes, the relation between wave height, depth and null-point grain size for waves of a steepness of 0.03 and 0.08 has been calculated (Fig. 41). The terminal fall velocities are according to the Subcommittee on Sedimentation (1957).

One finds that the often quoted opinion that there is no wave effect on the bottom below half the wave length (King 1959, p. 133) does not agree with the graphs. The null-point grain sizes are coarser than the sediments generally found on natural beaches off the break-point. The tank experiment showed the movement of material artificially distributed on the surface, but it did not show to what degree coarse particles at the origin situated near the shore-

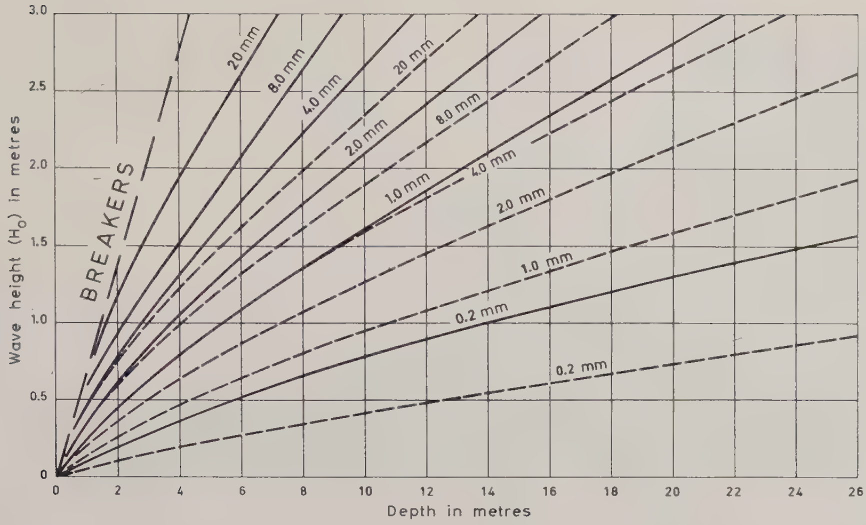


Fig. 41. Relationships between wave height, depth of null-point, and corresponding particle size according to formula (29). The solid lines refer to waves with a steepness of 0.08, and the dotted lines to a steepness of 0.03.

line can be transported seawards during a stage of rising waves. The quantitative meaning of the null-point is uncertain, because the effectivity of seaward transportation during increase of wave size (that is, during seaward displacement of the null-point) is not known.

In a review of the experiments King (1959, p. 133) makes the following statement. "In general it appears that the offshore motion on a natural beach will be very much less in volume and importance than the onshore movement and can be ignored for most purposes." This statement is difficult to understand. It would mean that the nearshore bottom is continuously supplied with material from deeper areas, whereas the reverse condition must exist, if any material is deposited off the surf zone. It is certainly significant that the experiments do not cover medium sand, fine sand and silt, which are frequently carried in suspension.

A straight bottom slope is hardly ever found in nature. The division of the movement on the experimental surface can be looked upon as an attempt of the waves to form a profile of equilibrium. From studies on Danish shores Bruun (1954, Part I) stated that in areas sufficiently fed or undernourished the profiles satisfied the equation

$$y^{3/2} = px \tag{31}$$

to a depth of $5H_0$. y is the water depth, x is the distance to the shore, and p is a constant that varies with the local conditions. Very steep waves are said to flatten the profile.

The formula (30) shows that, with increasing mass-transport velocities towards the shore, the angle of the slope and/or the sediment-particle diameters increase. According to the formula, the slope is steeper when the material is finer. Further experiments on the equilibrium profile with special respect to the boundary-layer conditions have been carried out by Eagleson and Dean (1959).

Depths of wave action

In Lake Vättern it is of special interest to know the depths of wave action, not only with regard to the present shores, but also for the determination of old submerged shore levels from the composition of the bottom sediments. The null-point of particles with a diameter of about 0.2 mm could be expected to give a fair estimate of the greatest depth of wave action. However, it has not been possible to find published data from natural shores, which prove or disprove with certainty the relation of formula (29). Our defective knowledge of the depths of wave action is well illustrated by King (1959). She cites the opinion of Cornaglia on a maximum depth of 10 m in the Mediterranean, a depth of 2.5 times the wave height according to Timmermans, a depth of about 3.5 times the wave height according to Inose and Shiraishi, small changes in depths greater than 30 ft at La Jolla, California, according to Inman and Rusnak (1956), but ripples later observed down to a depth of 170 ft (Inman 1957). Because of lack of knowledge as regards the wave conditions it is impossible to compare these statements and observations.

When the equilibrium grain size approaches the lowest limit of possible bed-load transportation, it also approaches the largest size which can be eroded. This occurs at depths at which the difference between the maximum shoreward and seaward velocities becomes very small. Even if this was not the case, the shoreward stroke (assuming a low slope) would throw the fine particles ($D < 0.15$ mm) into suspension, and they would thus also be moved by a slower seaward flow. At greater depths the sediment particles cannot be directly related to the wave motion. There the distribution depends on currents induced by wind and waves, which carry particles whirled up in shallower water, and it can also be influenced by gravity movements, such as slumping and turbidity currents.

An attempt to calculate the position of the greatest depth of wave action and the largest erodable grain sizes at different depths has been made by using Johnsen's formula of maximum shoreward velocities under wave crests (22). The formula has the advantage of being based on experiments in a big tank, which had a movable bed of sand with an initial slope of 1:50, which is of the same order as the slopes found in nature. The deviations of the measurements were small, but the most extreme peak of the maximum velocities may not have been recorded because of the type of velocity meter.

The application of the formula to the actual problem requires values for the critical erosion velocity. Values according to Sundborg (1956, Fig. 13) for a flow over a level bottom have been chosen (Table 12), because they are well established. The effect of the low bottom slope is thus neglected, but, on the other hand, the maximum velocities may be somewhat too low.

Table 12. Relation between particle size and critical erosion velocity. The figures apply to a uniform material of density 2.65 g/cm^3 and flow velocities 5 cm above the bottom.

D mm	V_{Crit} cm/sec
0.2	28
1.0	40
2.0	55
4.0	75
8.0	100
20.0	130

It has been assumed that no transport is initiated at velocities lower than 15 cm/sec. This agrees with Johnsen's observations (1961, p. 30) and experiments carried out in the flume of the Institute of Geography in Uppsala by Sundborg. The calculated relations between the largest depths of wave effect on different particle sizes of waves with a steepness of 0.03 and 0.08 are shown in Fig. 42. The diagrams can be compared with those of the null-point according to the formula of Ippen and Eagleson. If, firstly, the waves with a steepness of 0.08 are compared, we find that for a wave height of 0.5 m the null-point of 0.2 mm particles lies 32 % deeper than the greatest depth of initiation of transportation, and for a wave height of 1.5 m it lies 39 % deeper. For a particle of 1 mm the curves intersect for a wave height of 1.5 m, for lower waves the null-point lies at a somewhat greater depth, and for higher waves it lies at a smaller depth. The precentual difference for a 3 m wave is 12 %. The curves of coarser particles intersect in the same way. Near the break-point smaller wave heights are thus required, according to the null-point formula.

For waves with a steepness of 0.03 the differences are larger. The null-point curve of 1.0 mm lies close to the lowest limit of transportation. The 2.0 mm null-point curve coincides with the 1 mm curve of transportation, and in the same way the 4.0 and 2.0 mm curves lie close together. The curves which correspond to the 20 mm particles differ considerably–for a wave height of 2 m the null-point lies at 8 m and the initiation of transport at 4 m.

In this comparison it should be remembered that the diagrams have been calculated by very different methods, and that they do not apply to the same slope. However, it may be noted that for coarse sand and granules (the

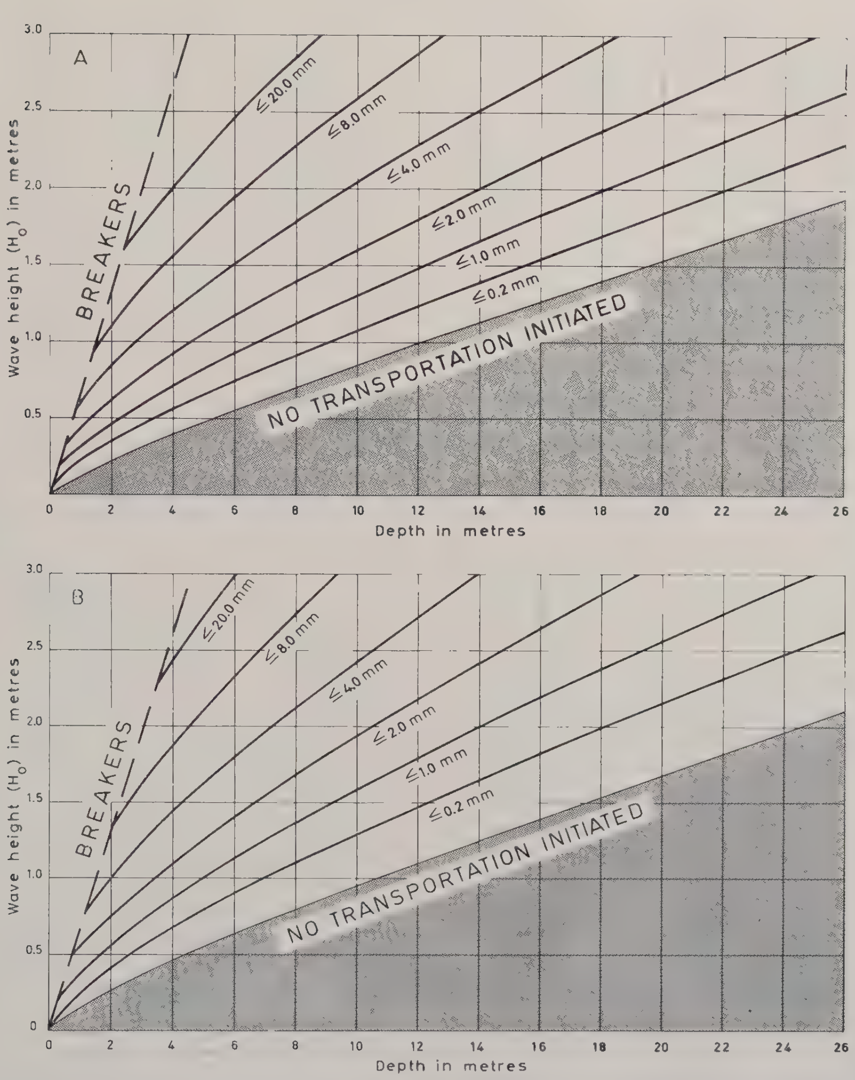


Fig. 42. Depths of initiation of transportation in relation to wave height for waves with a steepness (H_0/L_0) of 0.08 (*A*) and 0.03 (*B*).

particle sizes used in the experiment on the null-point) the differences are quite small. The coarser particles corresponding to the null-point near the break-point may be explained by the smaller loss of energy on the steeper slope.

The null-point of 0.2 mm particles lies at depths of 75 % to 85 % of the wavelength ($H_0 = 0.5$ to 1.0 m) for a wave steepness of 0.03 and at about 91 % to 116 % for a steepness of 0.08, whereas the extreme limit of initiation of transport lies at about 32 % to 36 % ($H_0/L_0 = 0.03$) and 70 % to 85 % ($H_0/L_0 = 0.08$) of the wavelength. A null-point situated at a depth of 28 m for a 1 m wave with a steepness of 0.03 undoubtedly seems very deep. The problem of the greatest depth of wave effect will be further discussed in the section on the Vättern profiles.

Wave action in profile

Zonation

In shoaling water there is a continuous deformation of the waves until they break. Shorewards of the common break-point the wave motion is characterized by breakers and reformed waves. The macroturbulence is high in relation to the velocities. The breakers end up in the swash of turbulent mass transport. Thus there are three discontinuity points, one where the waves start to initiate a transport, one at the break-point, and one at the lower limit of the swash. These points divide the profile into five zones of different transport and settling conditions, which correspond to a sedimentological and morphological differentiation: the deep-water zone, the shoaling zone (off the break-point), the breaker zone, the swash zone and the subaerial zone.

For a fixed profile the lower limit of the second and third zones moves seaward with increasing wave dimensions, and the swash zone broadens both seawards and landwards. The most diffuse limit is that between deep water and shoaling water. The lower limit of the swash is the most distinct, especially where there is a step.

When the bottom is steep and the waves are small, the waves do not break until they reach the step. A general zonation based on this condition has been proposed by Miller and Zeigler (1958), and they divide the profile into the zone of shoaling waves, the breaker zone (at the step), and the foreshore. However, this zonation does not cover the conditions of highest morphological activity, when the waves start to break far from the step.

In the deep-water zone there is only transport and deposition of fine particles eroded in shallower areas, as long as there are no strong tidal currents, slumping activity or turbidity currents. This sedimentological environment resembles the bottom-set bed of deltas. The character of the sediments largely depends on the organic production and the water chemistry. The sediments are often gelatinous and have a low shear-strength, which entails the possibility of gravitational movements at extremely low slope angles (1°–2° according to Arkhanguelsky, cf. Potter and Pettijohn 1963, p. 158).

The conditions of wave transportation in the shoaling zone off the breakers has already been discussed in the sections on the modes of transportation and the null-point hypothesis. This zone can be compared with the delta front. If an initial stage with a steep slope is assumed and material is produced near the shoreline by abrasion, the material moving seaward will form a slope off the abrasion platform. If the material is coarse (bed-load transport), the slope may be distinct and have an angle with the horizontal of 20–40°. This is a classical profile commonly illustrated in textbooks (cf. Ahlmann 1914, Johnson 1919), but not very common in nature. The shallow-water area is broadened by erosion at the shoreline and the accumulation of foreset beds. If a predominant part of the material which moves seawards is carried in suspension (which can be expected from the null-point hypothesis), the slope will be low.

Destructive and constructive waves

Although genetically not the whole truth, it can be said that the supply of material from the swash zone determines the morphology of the breaker zone. Waves which cause a loss of material on the shore face through seaward transport have been called "destructive" waves in contrast to "constructive" waves, which cause accretion on the shore face. The change can be regarded as attempts to adjust the profile to the wave motion.

It is commonly accepted that the constructive and destructive action can be related to the steepness of the waves. Lewis (1931), one of the first to describe this action as observed in nature, mentions that already in 1834 Palmer had noted the connection between destructive action and short-period waves. Lewis had observed the wave action on a shingle beach at Hurst Castle spit. The erosion or accretion on the shore face naturally depends on the balance between uprush and backwash, and Lewis describes the erosion as caused by the weakness of the uprush in comparison with the backwash (op. cit., p. 130). He also noted that the destructive action steepened the foreshore slope. A weak uprush and a strong backwash seems paradoxical, because the energy of the backwash depends on the uprush, and the relation is not clearly explained by Lewis.

As mentioned in a previous section, steep waves tend to form plunging breakers, in which much of the directed kinetic energy is transformed into confused turbulence. In the swash the plunging breaker curls over, and the crest falls down, either on the outflowing backwash or directly on the exposed shore face. I have examined this motion, while anchored to the bottom of the outer swash zone, where visual observations were made on the movement of air bubbles, solid particles and injected fluorescine solution. It seems as if the start of the plunging breaker up the slope takes place on outflowing water or on unspecifically directed vortices. In the upper part of the shore face vortices are continuously developed, so that near the bed the water mass consists of vortex rolls. This means that, although a water mass is transported up the beach, the motion near the bed is not effectively directed. Before the backwash starts, most of the vortices have disappeared, and in the flow down the slope the energy appears to be much more effectively directed.

There were no opportunities to study large spilling breakers in the swash zone, but it is evident that in general much less energy is lost by turbulence in spilling breakers than in plunging breakers.

Wave size, wave steepness, sediment-particle dimensions and slope are all related, and thus the critical steepness is not always the same (King 1959, p. 126). According to the Beach Erosion Board (1961, p. 83 *a*), laboratory studies indicate that the critical wave steepness is "in the order of $H_0/L_0 = 0.025$. This has not yet been confirmed in nature." From tank experiments King reports a variation of the critical steepness between 0.012 and 0.03. She also mentions an observation of the destructive action on a natural shingle beach of waves with a steepness of 0.019 (1959, p. 126 and 280). It will be noted that, although the estimates of the critical wave steepness vary, it is lower than the steepness generally apparent in Lake Vättern.

Shingle beaches form steeper slopes than sand beaches because of the increased rate of percolation of well-sorted shingle (Emery 1955). This is of importance in formulas on the stability of artificial rock fill. In such structures destructive action both from uprush and backwash has to be taken into account (cf. Hedar 1960 *c*). The slope of fine-sand beaches is very low but rapidly increases for medium sand.

On Californian beaches the slope corresponding to median diameters of 0.17 to 0.22 mm varies from 1:90 to 1:70, for a diameter of 0.35 mm it is 1:13, and for 0.85 mm it is 1:5 (Bascom 1951). On shingle beaches the gradient may rise to 1:2 (King 1959, p. 322), and Guilcher (1958, p. 80) mentions observations of the exceptional gradient 1:1.66 (31°).

The slope of the foreshore becomes steeper as the material becomes larger, as the wave length decreases, and as the wave steepness decreases (King 1959, p. 327). The steepening effect of decreasing wave length has been proved by tank experiments. The well-known fact that lake beaches are generally steeper than sea beaches may thus be explained by a greater effect of wave length than of wave steepness.

Sediment distribution

On ocean sand beaches the range of grain sizes of the foreshore is generally limited. The slope is slightly concave and its foot is often marked by a runnel, and there may also be a slight step (cf. Guilcher 1958, pp. 78–79). By a comparison of the size distribution across the beaches in five localities (the relative grain sizes in relation to the median at the mid-tide level are compared), Bascom (1951) proves the existence of a uniform type of variation. The sediment is coarsest at the plunge point seaward of the mid-tide level. Offshore from the plunge point, the particle size rapidly falls off, and then the curve of decrease flattens out. Shoreward of the plunge point the size decreases at about the same fast rate to the mid-tide level. From this point the rate of decrease is greatly reduced (there may even be an increase), and in relation to the general trend there is a secondary maximum at the top of the summer berm. From that the particles become smaller towards the dunes.

King (1959, p. 163) states that it is a well-known fact that, on a beach of mixed sand and shingle, the shingle is normally found at the top of the beach, and she refers to agreement with a tank experiment on particles between 0.5 and 0.9 cm by Bagnold (1940). This sorting is explained "by the fact that shingle is only in suspension actually at the break-point and therefore progresses by rolling on the bottom, where it is influenced by the differential velocity moving landward under the greater acceleration in this direction under the wave crests". The following remarks are quoted from her further discussion on the distribution on sand beaches (op. cit., p. 164). "In considering the variation of sand size on the profile at any one time it is found that the finer sand is on the upper foreshore, with coarser sand on the lower foreshore and on the berm crest. The lower foreshore clearly receives more energy than the upper foreshore, which is at the limit of wave action, but it is less easy to explain the coarse sand on the berm crest. This may partly be due to the winnowing effect of the wind, as the berm crest is only reached by very high water on rare occasions." The explanation used on the "mixed" beach is not applied. The problem of this sorting will be further discussed in relation to the Vättern shores.

Because the step, except on some shingle beaches, is a small and not very significant feature on ocean beaches, it is generally not paid much attention. However, on lake shores, where the material is generally not so uniform, the step is perhaps the most significant feature of the foreshore. A theoretical model presented by Miller and Zeigler (1958) is therefore of great interest. In this model the sediment pattern is related to the dynamics in the zone of shoaling wave, breaker, and foreshore, and the breaker zone is concentrated to a well-developed step. The theoretical considerations are primarily based on the work done at the Massachussetts Institute of Technology previously referred to in the section on the null-point hypothesis.

As regards the sediment pattern of the model, the particles are coarsest on the step and become smaller in the shoreward direction in the zone of shoaling waves, and also in a landward direction on the foreshore above the step. In the shoaling zone the sorting is increased towards the breaker zone (the step), but close to it expected to be quite irregular. The sorting is best on the step, and then gradually decreases up the slope of the foreshore. The authors found this model to be in "sufficiently good" agreement with available data from natural beaches (Falmouth, Highland and Lake Michigan Beach).

As regards the dynamics, the hypothesis of net vertical movement in the base of the turbulent breaker with a secondary sorting action on the material (op. cit., p. 547) is most noteworthy. The hypothesis is supported by sampling in the breaker itself, which showed a distinct gradation in a vertical direction, from coarse and large volume at the bottom to fine and least volume at the top. This agrees well with my own visual observations of vertical movement over the step. By this action the fine particles are separated from the step sediment.

The model in general agrees with the sediment pattern described by Bascom and King, except for the top of the berm of sand beaches, and the foreshore distribution of shingle according to King. The pattern of sorting seems not to agree with a study by Inman (1953), in which the sorting in the surf zone was found to be the poorest. However, the sorting over a broad surf zone is not comparable with the sorting on a distinct step.

There are several questions with regard to comparisons between different studies. Two of the most important are: under what conditions was the material sampled, and can the sediment be related to specific dynamic conditions? Undoubtedly most samples are taken during periods of rather calm weather, that is to say, the sediment pattern probably reflects the last period of wave decay and an unknown number of hard-weather periods. The material at the top of the berm has not been transported by the same type of waves as has determined the surface sediment on the lower part of the foreshore. The sediment corresponding to the top of the berm is either buried or eroded. If there was a deposition during the wave decay, finer material can be expected on the lower parts of the foreshore. An explanation of the pattern in detail requires information about the wave history, if samples are taken above the zone of actual wave action during the sampling.

Form elements of the nearshore zone

The morphology of the nearshore zone off the foreshore has been discussed in great detail by King (1959) from the results of her own and other experiments and data from natural beaches. In this paper it seems convenient first to give the important conclusions from her work and then to discuss some investigations on wave action and nearshore morphology not referred to by her.

In King's experiments it was found that a bar forms at the break-point of the waves, where the sand moving landward outside the break-point accumulates along with the sand moving seaward inside the break-point of steep waves. The conditions necessary for the formation of this break-point bar were found to be linked with the relations between wave steepness and bottom gradient. The relationship is shown to be fairly linear. For a gradient of 1:20 a bar can be formed, when the wave steepness exceeds 0.0115, but for a gradient of 1:5 the steepness must exceed 0.034. A second bar, landward of the first, was developed, if the wave size was suddenly made smaller. Increase of the waves moved the material of this bar shoreward to the first bar. The break-point bar was destroyed by a falling water level.

Another type of bar was formed, when the gradients were low and the steepness small. It forms in front of the break-point of waves of low steepness, if the original gradient is less than the gradient built up by the swash of the waves in their constructive action. There is a deepening of the water outside the break-point, which shows the source of the sand built into the swash bar. The height of the bar crest is raised above the still-water level to the limit of the swash. King stresses that the break-point bar and the swash bar are two distinct forms, which cannot develop the one from the other (op. cit., p. 191).

The bars of the experiments are identified with forms found in nature. The break-point bar is identified with submarine bars, also called longshore bars. [King (op. cit., p. 331) lists papers on these bars, from the first by de Beaumont (1845) to the modern ones by Evans (1940), Keulegan (1945, 1948), Shepard (1950), King and Williams (1949, 1951) and Bascom (1953). The following may be added: Vollbrecht (1954, 1957), Körner (1955), Johnsen (1961) and Davidsson (1963).] The submarine bars lie parallel to the shore. They are generally three but sometimes two in number. The outer bar is said to be formed during very rare and extremely stormy conditions, and is not affected by the ordinary waves, which break nearer inshore and form the middle and inner bars. The last is the most mobile, and moves seaward with increasing wave heights and landward with slowly decreasing heights. The submarine form is typical of tideless coasts. It is essentially a destructive form connected with the break-point and thus cannot be built up above the water level (King, op. cit., p. 337).

As regards the terminology, longshore bar is more neutral than submarine bar with respect to limnic and marine conditions, but in my opinion the exclusive formation below the water level is most important, and in lakes the term "subaquatic bar" can be used.

The swash bar is identified with a morphological form called ridges and runnels. Developed in tidal seas, the ridges will form most readily at the mean high and low tide levels. Like the experimental swash bar, they are a constructive form in front of the break-point, and are built up to satisfy equilibrium conditions on an initially too flat surface. The outer bar is built up during low tide, and then there may or may not be time enough to build up a bar at mean tide level, but again a bar is built up at the high tide level. The ridges and the runnels between them can be parallel to the shore or diverge from it, depending on the direction of the waves which form them. They can be expected to be most readily formed in areas with limited fetches, because the steeper waves in such areas require a steep equilibrium profile. King stresses the connection between tide and ridges and runnels, but it is also important to note her remarks about the favourable conditions of a "super-abundance" of sand (op cit., p. 342) found in flat areas of glacifluvial deposits, such as the former outwash plains of the Netherlands and Denmark and the sandurs of southern Iceland.

The large scaled forms of barrier beaches (barrier islands, offshore bars), which fringe the eastern coast of the United States, the Gulf of Mexico, the Icelandic sandurs, etc. are also included in the constructive forms identified with the formation of the swash bar (op. cit., p. 353). When the tidal range is small, the process can build up a barrier beach that is permanently above the mean sea-level. Small forms of the same type, comparable in size with the submarine bars, are not discussed.

It seems difficult to distinguish a swash bar (a form which also in nature could be called a swash bar) from a beach ridge, when the bar is built up in direct contact with a previously existing foreshore. The swash bar may be called an offshore bar when developed without contact with the shoreline, but the essential thing is the attempt to form a beach in equilibrium with a specific wave action on a bottom which is too flat to correspond to the position of the present shore. This bar could as well be called a beach bar. The existence of ridges and runnels submerged by the tide is, as it is genetically explained by King, a secondary phenomenon. They must be regarded as swash (beach) bars which survive in favourable conditions.

Schou (1945, pp. 68–69) describes an onshore movement of submarine bars, which under favourable conditions (high water, translatory waves) move up on the shore. The most shoreward bar is called the migrating bar. In lectures given in Uppsala in 1961 Schou declared that he was convinced that the submarine bar as a form cannot move into the position of a swash bar.

The term "migrating bar" has recently been taken up by Davidsson (1963). The littoral zone is illustrated by a fundamental sketch (op. cit., Fig. 2) with (1) a beach ridge, (2) a migrating bar with the crest below the mean water level (this is also said to be the area of sand banks) and (3) three longshore bars. In this sketch the migrating bar and the two first submarine (longshore) bars are about equally spaced and the outer bar lies somewhat more isolated, and it also looks more like a low terrace. The same relations are found in the true main profile of the investigation (op. cit., Fig. 23). Davidsson's investigation is of great interest, but unfortunately there is no conclusive basic discussion. The character of this migrating bar is unclear, and therefore scattered sentences have to be cited.

> The area between the 2. bar (the middle submarine bar) and the beach is characteristic of constant changes. The relation of morphology to wave work cannot be drawn up, as every wind situation results in a specific wave and current action and thereby a specific morphological work (p. 31).—The waves break on the migrating bar (p. 49). —The unstable migrating bar runs along the beach of the west side. The off-shore form can be looked upon as an off-shore barrier, and like this it appears at low water (p. 85). —The hard wind that blew between the 26th and 27th produced destructive waves with a thorough influence on the form of the breaker zone. The migrating bar was levelled and moved from the beach and that was also the case with the bars (p. 103).—The migrating bar received sand from the trough inside and from the beach. At the same time as the migrating bar was built up it also lost sand to the 1. bar. There entered sand into this also from the trough between the migrating bar and the 1. bar (p. 118).—In front of the original waterline lies a formation reminding of a migrating bar, which at high water has moved so far towards the beach that, after the average water-level has returned, it has remained as a kind of beach ridge. The original trough was nearly free from water but it had earlier been a lagoon (p. 114).

It should be mentioned that the last quotation refers to a levelled profile, and that the preceding levelling, when the migrating bar was identified just below the water level, took place seven days earlier.

From these sentences it can be questioned whether this migrating bar is a special form not taken into account by King. Is it a kind of swash bar or is it just the innermost submarine bar? The last alternative seems most probable. As far as can be judged from the descriptions, the bar is never built up above the water level, but it may accidentally come above the water level if this falls off. The gradient of Davidsson's main profile is about 1:100, which, according to King's experiments, results in the formation of submarine bars, even at an extremely low wave steepness. The position of the bar may be favoured by this and by an abundance of suitable material.

However, the position of the migrating bar in the same area as a form which Davidsson calls a sandbank (not defined) draws attention to the possibility that the migrating bar (or the sandbank) may in some cases be identical with a bar form built up by a predominant onshore swash in front of the shoreline. The water flowing over the bar moves along the shore until it flows out in concentrated currents (rip currents), called by Davidsson "cross currents" (op. cit., p. 33). The stratigraphy of this type of bar or bank resembles fluvial banks, as material is deposited at the angle of repose on the shoreward, distal part. A bar of this form is described by Schou as a migrating bar (1945, Fig. 15). It seems to exist in areas with an abundance of sand and an extremely low gradient. I have examined such bars in the northern part of the Gulf of Bothnia, on the front of the delta of the Indalsälven river, in Laholm Bay on the Swedish west coast, and in Karrebaeksminde Bay on south-western Zealand. The bar is not built up above the water level as long as the swash flowing over it can be drained off. This type should be called a swash bar, as it is built up by the onshore movement of particles under swash action. The migrating bars described by Davidsson can be judged to be submarine bars in a genetical sense, but in some cases their relation to swash action is uncertain.

A somewhat different swash bar, connected at one end with the shoreline and diverging from it, is described by Schou (1945, pp. 74–75). He refers to the term "cusplet", in-

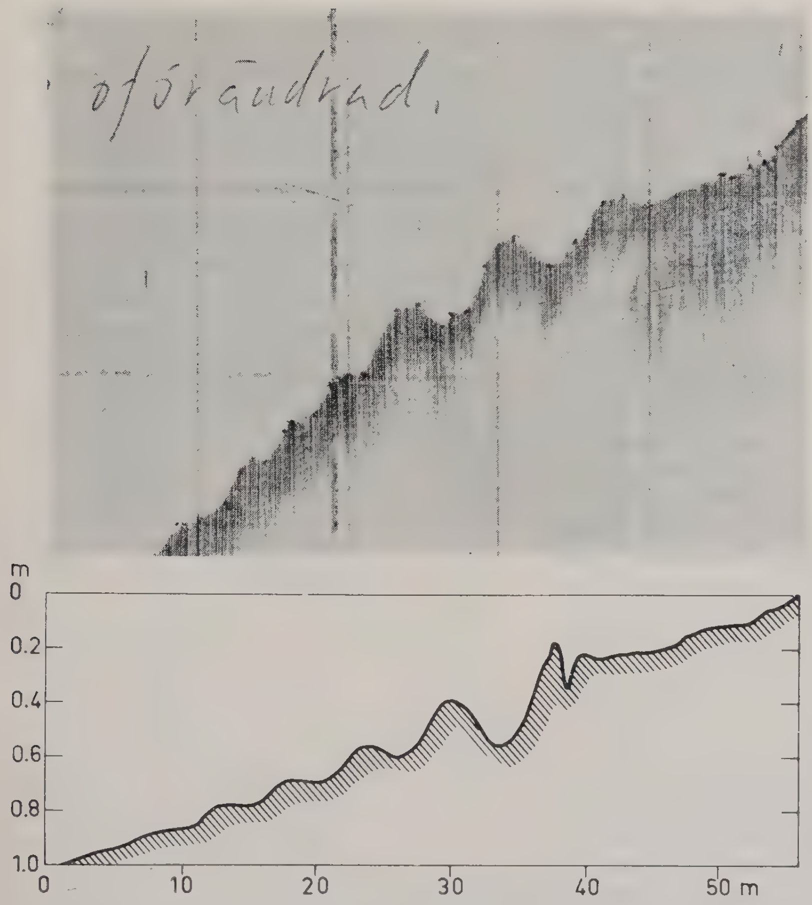


Fig. 43. Echogram of a profile with bar forms off Gränna harbour (cf. Fig. 4) and bars developed in a wave tank (from Johnsen 1961). Depth interval of the echogram 15–30 m. Black spots on the contour are pencil marks. The experimental bars were formed by waves with $H_0 = 21.9$ cm and $L_0 = 7.08$ m ($H_0/L_0 = 0.031$).

troduced by Wilson (1904), but calls it a diverging beach ridge. The diversion from the shore is explained as an adjustment to wave direction, which agrees with King's opinion on the orientation of swash bars. I have observed this type of bar in the northern Bothnian archipelago and on the shore of the Lule Älv river at Gammelstad, where the river is 3 km wide and the current of the main channel does not act on the shore. The gradients of the nearshore bottom were everywhere small. As in the previously described swash bar, this bar has a distinct distal slope. It is interesting to note that the bar is built up above the water at the connection with the shore, where a free swash over it is hindered and the material moved by the swash settles on top of the bar before a backwash starts.

A theory by Johnsen on the genesis of submarine bars (1961) will finally be discussed. His work is based on experiments on the effect of differential velocities under crests and troughs of waves running up on a shore. Some of his results with regard to the wave motion have already been discussed in the section on the null-point hypothesis.

The initial bottom slope in the experiments was constantly 1:51, and waves with a steepness varying between 0.031 and 0.042 were generated. Thus the experiments only covered waves of a destructive character. The sand had a median particle diameter of 0.16 mm. One important result immediately found from the experiments is that in all trials with a uniform wave size more than one bar was formed. One of the profiles is reproduced in Fig. 43.

The explanation of the genesis of the bars is founded on the movement of material above ripples, i.e. on a small form dependent on the wave motion. According to Johnsen, the particle volume moving over the ripples can be divided into three parts, which are dependent on oscillatory components: (1) the contact load moved from the proximal to the distal slope of the ripple, (2) a part which is lifted in suspension directly in the direction of the flow above the ripples, and (3) a part which moves through the leeward vortex and is turned by the contrary oscillatory component. The two first parts become one because they have the same direction and consist of the coarser material moved.

On the basis of the distribution of the oscillatory components of the wave motion over the nearshore bottom, Johnsen constructs (qualitatively) "mass lines" which, when added, form balance (Überschutz) lines of the particle movement. These lines of the balance of the movement in seaward and shoreward directions give qualitative information about the net motion of the different parts of the bottom, and they predict the relief. Two main cases of particle movement are distinguished with regard to the relation between wave height and the median particle diameter. In the first case there are large waves passing relatively small particles. Then the particle movement is said to be mainly dependent on the part, which moves through the leeward vortex, which results in a seaward balance. In the second case low waves meet relatively large particles. The directly transported part is the part which moves through the leeward Between these two extremes there is a transitional case, which also corresponds to the relative minimum quotient between wave height and particle size at which bar formation is possible. With a smaller quotient only a sandbank can be formed. The upper limit above which bar formation is impossible corresponds to a state, when the main volume of the particles taking part in the movement is suspended throughout the wave period. The two main cases and the transitional case are illustrated in Fig. 44. It can be seen that the break-point is moved with the formation of a bar. The formation of the outer bars can be explained as an effect of the first bar formed, and, according to Johnsen, outer bars can only be formed without direct influence of a breaker, when the net movement is seaward. It may again be remarked that in none of the experiments was a single bar formed.

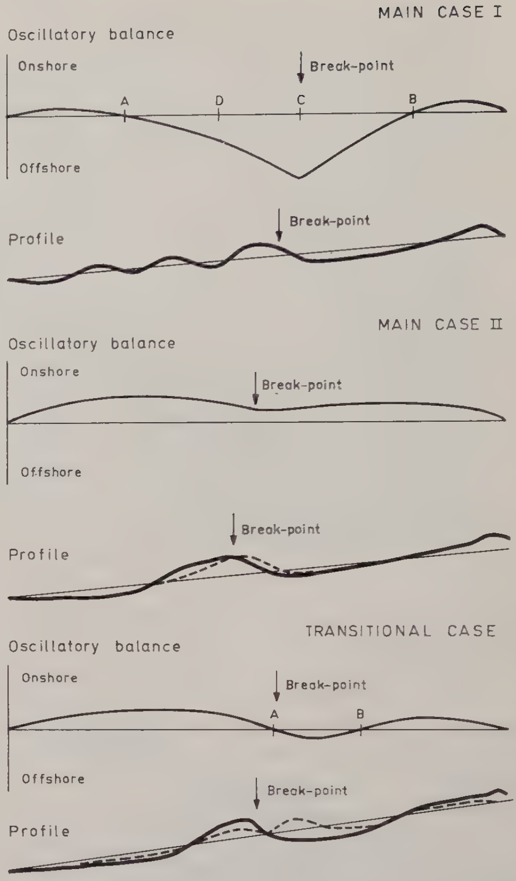


Fig. 44. The two main cases of subaquatic bar formation and the transisional case in between them. For further explanation, see the text (from Johnsen 1961).

The wave size and the particle distribution also affect the depositional direction and the bar formation. The wave type is said to be of less importance as regards the introductory formation. The bars are almost only built up of material transported seaward from the area close to the foreshore. It may be noted that, as far as uniform material is concerned, only the balance of the transitional case is in conformity with the null-point hypothesis.

Johnsen stresses that the bar is a formation of equilibrium character in contrast to sandbanks, and he discusses the different conditions of equilibrium in relation to transport balance and particle size. Changes of waves and water stage cause changes in the form and position of the bars, especially the most shoreward one. Increasing waves cause a seaward transport, and the bars also move seawards. A decay gives a reverse transport and change in position. A rising water level without change of the waves causes a landward transport off the break-point and inside a seaward transport. Thus the bar moves landwards. With a falling water level the change will be the reverse.

The investigation reported is in part qualitative, and the waves were paddle-generated (an onshore wind can be expected to increase the seaward transport (cf. King 1959, p. 213)), but the results are most encouraging, and it is hoped that it will be possible to test them in nature by field investigations which have started this winter (1963–64) on the Outer Banks in North Carolina and will be carried out by the Coastal Studies Institute at Louisiana State University.

Conclusion

It is obvious that the depositional form which can be expected in the nearshore zone of lakes with their steep waves is the subaquatic bar. The possibility of development of swash bars can be judged to be small, except in very shallow and flat areas. From Johnsen's experiments it seems to be certain that the outer submarine bar does not necessarily correspond to the position of a break-point in extreme storm conditions.

The maximum depth of the direct influence of the wave motion cannot be predicted with any certainty. Opinions on this depth are too divergent.

The wave action in the swash zone can be expected to be of a predominantly destructive character, involving abrasion and formation of steep beaches with greatly differentiated particle distributions. The coarse material will form a step at the base of the foreshore.

Wave action in plan

Longshore drift

The lateral action depends on the direct effect of oblique wave motion and on the wave-induced currents. The longshore movement is most important, because it is responsible for the displacement of much larger quantities of material than the movement normal to the shore. It is also responsible for the development of the shoreline configuration, both with regard to erosional and depositional forms.

The main part of longshore movement takes place in the swash zone and in the breaker zone. By the oblique uprush and more perpendicular downwash the particles in the swash zone move in the well-known zig-zag transport or beach-drifting along the shore. In the breaker zone the particles move to and fro under the waves, but are laterally displaced from the net direction of this movement by even the slowest longshore current, especially when thrown into suspension by the breakers. In the inner breaker zone the longshore current may have the character of a continuous flow capable of transporting pebbles.

The interrelation between waves and currents and the variation in grain size make formulas of general relations between longshore currents and drift volumes suspect. The formulas relating wave energy to drift, based on investigations on Californian sand beaches and model experiments (summarized by Savage 1959 and the Beach Erosion Board 1961), have already been discussed in connection with the currents. Here only some conclusions from a model study by Saville will be added. It was found that transport under steep waves mainly takes place in the breaker zone, while transport under low waves takes place predominantly in the swash zone. The critical steepness of this balance was 0.03, and the maximum rate of transport was associated with a steepness of 0.025 (Saville 1950). It may be noted that the wave-steepness interval (0.02–0.03) is critical with regard to the constructive and destructive action on the foreshore, the formation of submarine bars and the longshore drift. It has not yet been possible to integrate these partial relations into a general quantitative relationship.

By the longshore drift the beach and nearshore sediments will be laterally graded. A general decrease in grain size can be expected from areas of high energy towards areas of low energy, a gradation which is also found in nature, when, for example, exposed headlands are compared with sheltered bays. The greatest variation will be found, where only residual material is left in the most exposed areas.

Along quite straight shores the direction of net movement depends on the balance of the azimuthal wave-energy distribution. Because of the different effects of low and steep waves and the different resistances to erosion with regard to particle size, the movement on the beach and in the breaker zone and the movement of different particles are not necessarily in the same net direction. One example of this is the yet unsolved problem of the development of Benacre Ness in Suffolk (Williams 1956). The possibility of different net movements on the beach can be theoretically exemplified by a shore of shingle and sand, affected by a wave-energy distribution, which consists of a high frequency of low-energy waves, giving one net direction of movement, and a low frequency of high-energy waves, giving the reverse direction. If the low-energy waves move the sand but cannot move the shingle, and the high-energy waves move both sand and shingle, but a smaller total volume of sand than the low waves because of their low frequency, the result will be a separation by different directions of net movement.

Form elements

A series of depositional forms generated by beach-drifting has long since been recognized and described (cf. Gilbert 1885, Johnson 1919, Schou 1945, Steers 1948, Cotton 1946, Guilcher 1958, King 1959). The material moved into bays forms *bay-head beaches,* which may broaden to a series of *beach ridges*. Where there is a sharp convex bend in the shoreline, the material tends to move on in the direction of the transport shore, but because the beach-drifting is bound to the swash zone and because of refraction (Evans 1942), the end of the *spit* built out will be recurved in the way so eminently exemplified from the Danish coasts by Shou (1945). Unidirectionally built *recurved spits* are formed at the end of forelands with a definite predominant drift direction. They are especially well developed in narrow inlets and bays, where the leeward activity is limited by the fetch. Shepard (1952) has proposed the term *barrier* for the types of accumulations which are separated from the shore by a lagoon. Synonymously a ridge enclosing a bay should be called a *bay-mouth barrier* (the term "bar" should be restricted to structures which are permanently under water (King 1959, p. 255)).

On headlands, where spits are built out from two directions, a converging system of spits forms a *cuspate foreland,* generally of triangular shape. The necessary qualification for the formation of a stable foreland is the balance of drift between the two flanks. With regard to the coast configuration there are two extremes. One is a narrow island with the wave energy almost solely distributed in one direction along its long axis. In this case a narrow spit will develop in the downdrift end like a downstream bank in a river. The other extreme is a rather straight or even indented coastline, where the foreland is built out by two wave-energy maxima of oblique and contrary incidence to the coast. The qualification of two distinct oblique maxima can generally only be fulfilled by the distribution of fetch, when, for example, the shore is situated on a narrow lake or in a straight. Dungeness on the English Channel is the most famous foreland of this type (Lewis 1932, 1940). Where the coast is sheltered by an offshore island, the foreland forms the embryo of a *tombolo* (Schou 1945, p. 133), the feature that will connect the mainland with the island.

There can only be a retreat of the shore, where the beach has been eroded, and in downdrift areas there cannot be a severe retreat, because beach-drifted material cannot be effectively transported seawards. These statements are very simple and not new, but they are important because they link the shore retreat to the processes of longshore movement. Almost all money spent on shore protection goes on the construction of groins and jetties, which, it is hoped, will collect beach material and prevent deficits in longshore beach-drifting. (It is questionable if they are effective in the long run in areas of submergence (cf. Bruun 1962).)

Shoreline configuration

Which shoreline configurations does wave action give rise to? The depositional forms mentioned are only a partial answer. The question can be divided into two parts, which both apply to retreating (erosional) and advancing (depositional) shores. One part refers to the geometrical shape, the other to the orientation of its features.

Much attention has been paid to the depositional forms in morphological literature, but less to the orientation, though there are important works on this subject, especially by Lewis (1931, 1938) and Schou (1945, 1952), and least attention has been paid to the relation between the form and the orientation of retreating shores. The general cyclic ideas of Davis (1912) and Johnson (1919) should not be overlooked, but they are founded on too little knowledge of the factors which have governed the development of the shores. For example, the offshore barriers of the American east coast are Johnson's classical example of a low coast straightened during emergence, but this coast is certainly a coast of submergence (cf. Russell 1958, p. 596).

The morphology of retreating shores was a main subject of my first studies on Lake Vättern, exclusively dealing with Visingsö island (1959). Here the geometry of the shoreline configuration will first be discussed in principle, then the relations between exposure and orientation, and finally the configuration of the Vättern shores will be discussed in a special section.

The wave-energy level increases in the downdrift direction of a straight beach because of increasing fetch. Tanner (1962, p. 39) states that, if the drift material was "whisked away by pushbutton magic", eroding drift would make the shore "look something like one-half of a parabola", with the vortex located at the point where the waves first become effective. But he finds that, because the material is not whisked away and shores are not infinitely long nor perfectly straight, a "curve of deposition" must be added to the "curve of erosion". "The amount of material removed in the updrift concavity must be equalled, when calm conditions return, by the amount deposited farther down the beach. Therefore, even on a straight coast, the updrift end of the beach should undergo erosion. Farther down the beach net deposition should occur. The result should be rotation of the strandline about a "fulcrum" somewhere between the two stretches. The new orientation of the beach will be a bit closer to the perpendicular to unrefracted wave attack."

As regards the position of the point of maximum net erosion, Tanner finds that this depends on (*a*) the linear distance over which the drift is moved during an average period of transport, (*b*) the ratio between the fetch at the downdrift end and the fetch at the updrift end, (*c*) the long-term uninterrupted duration of the process, and (*d*) the average energy level of the coast. The maximum point moves down from the updrift end with increase of (*a*), (*c*) and (*d*), and decrease of (*b*).

Tanner does not discuss the effect of reverse drift and the effect of loss of material. In many areas of severe erosion little of the material eroded has a grain size coarse enough to be kept on the beach, and even coarse material is lost to the nearshore zone, especially in areas of submergence (cf. Bruun 1962). Thus a certain "pushbutton magic" exists, and an erosional indentation skewed towards the updrift end–as between groins–can be expected. Where there is some beach material, this will be found at the downdrift end. With a reverse drift direction the points of maximum erosion and deposition will change places.

If the effects of the contrary drifts are fairly balanced, a symmetrical indentation will form with a smaller radius of curvature at the ends than in the middle part. If the drift in one direction predominates, the form will be skewed.

Geometrical forms of equilibrium of coasts with a littoral drift have been proposed by Bruun (1946, 1954), who states that the shoreline configuration can be described by a cycloid curve, the diameter of the corresponding circle being $A/2k$, where A is the wave effect and k is a constant (1954, p. 6). The formula is applied both to bays and headlands (islands). The headlands thus have an equilibrium configuration which is the reflected image of a bay. This convex cycloid is the same as is found along the front of a recurved spit.

The theoretical foundation of Bruun's equilibrium form has been criticized with regard to erosion and deposition in relation to wave incidence along the cycloid curve (Rüdinger 1955).

Bruun has exemplified his thesis by the fit of cycloid curves to the downdrift ends of some islands and of bays in the Danish archipelago. Nowhere can the fit be said to be exact, and it generally seems to be better in the bays. The deviations can be explained by lack of homogeneity in the natural deposits. However, a study of the configuration of the islands and headlands of this archipelago (which can be made on the topographic maps on the scale of 1:100,000 or 1:25,000) makes it clear that the smoothed convex shape is not the typical one. Instead we find that, although islands and headlands, broadly speaking, have a convex form, the shoreline is made up of a series of shallow indentations, which on the map look like garlands hung on a wall. These "garlands" are certainly, as a rule, "hung" on more resistant parts of the coast, which are not taken into consideration in a theoretical model of homogeneous conditions. This homogeneity makes possible a study of fundamental relations, but on the other hand, differences in erosional resistance are the rule in nature, and are therefore of significant importance on retreating shores. The function of the more resistant parts resembles that of the groins on the beach; only the existence of beach material can prevent a retreat.

Tanner (1962) demonstrates that, on the convex lagoon side of barriers, an equilibrium form consisting of a series of bays and spits can be developed. The configuration is characterized as having a "scalloped outline". This is expressive, and here the term *scalloped headland* is proposed (in the Visingsö study I used the term undulatory headland).

Shoreline orientation

The rule that beaches tend to orientate themselves perpendicularly to the dominant waves is ascribed to Lewis (Guilcher 1958, p. 182). The concept of "dominant waves" is not quantitatively defined, but is also described as the most powerful waves (Lewis 1931, p. 130). It should be stressed that the object of Lewis's investigations was shingle beaches, and especially beach ridges, and that the beaches studied were not generally straight but arch-shaped. It may also be noted that he did not investigate the nearshore bottom morphology. Of importance is Lewis's statement that "There need be no definite connection between the direction of prevalent waves which cause the drift of shingle along a coastline and the dominant, or larger storm, waves which I have suggested influence the direction of ridge building" (op. cit., p. 147).

The idea that from the wind distribution a directional resultant of the wind work (DRW) can be constructed is due to Schou (1945), and the direction of the DRW is assumed to represent the direction of the "dominant waves" in Lewis's sense. Schou's method cannot be regarded as being founded on the dynamics of wave action, but on the movement of particles of unit grain size over a homogeneous surface, the net movement caused by a distribution of forces acting from different directions. The method is based on the primary assumption that the effect of wind work (A) for a certain direction can be expressed by

$$A = k \sum_{B_i > 4B} p_i B_i, \qquad (32)$$

where k is a constant which, applied to the DRW, can be taken as 1, and p_i is the percentage of the wind velocity, measured on the Beaufort scale (B_i). The values of A for all wind directions are added by vector summation, and the resultant is the DRW (cf. Fig. 45 A). It may be noted that Schou sums the frequencies of winds stronger than 4 Beaufort. This makes it questionable if the DRW represents the dominant waves or the prevalent waves. Therefore directional resultants based only on the hard winds ($\geqq 7\ B$) have also been used (Davidsson 1956, 1963), but no specific lower limit can be stated to be correct in general. For example, on inland waters the winds stronger than 7 Beaufort may be so infrequent that it is impossible that they should determine the shoreline evolution.

Other formulas of wave force, represented by the wind velocity, have been applied to Schou's method by Christiansen (1957). The formulas applied, proposed by Munch-Petersen (1918), Bruun (1954) and Lewis, all take the fetch length into account, their general form being

$$A = F^m \Sigma p_i B_i^n, \qquad (33)$$

where F is the fetch length. In the formulas compared by Christiansen, the exponent (n) of B_i varied between 2 and 15/4, and the exponent (m) of F between 1 and 5/4. The exponent 15/4 in Bruun's formula is based on the wave height as proportional to the 5/2 power of the wind velocity in m/sec and the relation $B_i^{1.5} = U$ (m/sec).

The effect of waves generated by winds, represented by vectors parallel or almost parallel to the coast, is difficult to estimate, as it greatly depends on the refraction. Munch-Petersen (1918, p. 205) proposed a general correction by giving such vectors a 50 % weighting.

It has been shown (Schou 1945, Landsberg 1956) that the DRW can be correlated to the morphological shape and the drift direction of blown sand. This could also be expected, if the relation in (32) is not too far from reality, as the conditions in sand dunes are fairly homogeneous. It should also be noted that, if the exponent of B_i is not entirely correct, this has little effect upon the resultant, as long as the wind-velocity distributions in all directions are of the same type.

The application of the method to coast and shoreline orientation is purely empirical. In the case of small islands, all wind directions are taken into account. When applied to other coasts, offshore winds are excluded. According to Schou (1945, p. 216), the following relations exist between the DRW, the maximum fetch length, and the coastal orientation:

> Where the fetch length is equal in all directions, the terminal direction of coastal simplification will be at right angles to the DRW on both receding and advancing shores, this direction giving the minimum of beach-drifting.
>
> Terminal directions may also occur parallel with the DRW. On the lee side of islands, it will thereby be possible for a marine foreland to develop, with its longitudinal axis parallel to the DRW.
>
> If the maximum fetch length and the DRW coincide, similar conditions will prevail.
>
> Where the maximum fetch length and the DRW have diverging directions, the compass direction governing the terminal direction in the above manner will lie in the angle between the maximum fetch length and the DRW.

The main dynamic interpretation that can be made of these statements is that, in the case of a continuous coast, the orientation will tend towards a state of least beach-drifting, and then the drift in opposite directions will be in balance.

In the case of a coast with homogeneous conditions as regards sources and probability of transportation, and in a state of retreat through biased beach-drifting, the general tendency to re-orientation proposed by Schou could be questioned (cf. Bruun 1946). It could also be questioned whether the DRW is synonymous under all circumstances with the "dominant waves", this concept properly being attributed to one direction, whereas the DRW may be the product of several directions, no single one being predominant.

I have examined in detail all the examples given by Schou, and found it difficult to see that these examples really prove the DRW rule. Davidsson (1963) has tried five different types of directional resultants, his own being based on calculated wave energies, and the result is not very positive. "Regardless of which of the 5 methods that is used, it is obvious that the only function of a resultant is a complement to other indicators of resulting beach-drifting" (op. cit., p. 97).

Only one of Schou's localities will be discussed here, the island of Anholt, situated in the Kattegatt between Denmark and Sweden. The configuration of the island and Schou's DRW, based on all wind directions, are shown in Fig. 45 A. According to Schou (1945, p. 58), the locality illustrates well the correlation between a DRW, based on all wind directions, and the development of shoreline configuration. The maximum fetch length has not been used by Schou in this locality, and he gives the explanation that in all directions the fetch is large enough to generate a strong wave effect. However, it should be noted that the fetch length of the resultant direction is 40 km, but the fetch length from the NNW. is ten times longer.

The different variations of the directional-resultant formula which have been previously mentioned have been applied on the island by Christiansen (1957). The resultants all lie within an interval of 18°. The maximum theoretical error of the vector construction as such is 22.5° (when based on eight wind directions), but, in reality, this error is generally much less, and in a comparison between different methods, applying the same wind distribution, it is of small importance.

As can be seen from Fig. 45, Schou's DRW deviates 20° from the longitudinal axis of the leeward accumulation, and, according to Christiansen (op. cit., Fig. 2), the resultant constructions, which take fetch into account, deviate more towards the south-east. This shows the effect of the long fetch from NNW. The conclusion drawn by Christiansen (op. cit., p. 36) that, because the different resultants lie within an interval less than the maximum error of a single construction, Schou's original method can be regarded as useful and has the advantage of being the simplest is not very satisfactory.

It has been suggested that in channels or other areas with two maxima of fetch length and wind force, the flanks of a projecting head or spit should be correlated to two resultants, obtained by splitting the wind directions into two parts (cf. Guilcher 1958, Fig. 27, which

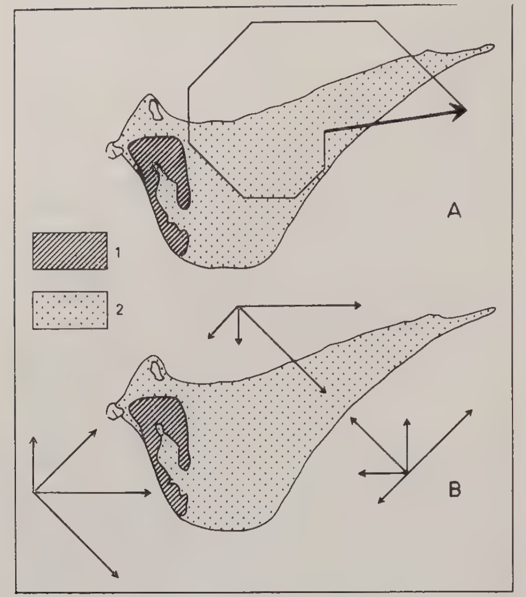


Fig. 45. The shore configuration of Anholt island. 1, kernel of moraine. 2, marine foreland. *A*, the DRW according to Schou (1945). *B*, the vectors of wind work which affect the three main shores.

exemplifies the conditions at Dungeness). This method can be applied to Anholt by a separate treatment of the components which have a direct influence on each of the three sides of the triangular island.

The three vector groups obtained from Shou's data are drawn in Fig. 45 B. The western group indicates a main transportation around the southern part of the island, a transport direction which nowadays is reinforced by the harbour built out from the western point. The two other groups indicate a drift towards the eastern spit; the northern group, being more askew, could be supposed to reflect a higher potential drift. It is important to note that both the shores of the accumulation area have the shape of well-defined arcs the southern being almost symmetrical and the northern askew, and that this is in conformity with the vectorial distributions. In my opinion the eastern part of Anholt can be regarded as a foreland built out from a coast. That this foreland is not linked to a coast, but that the material is taken from the western shore of the island is of less importance.

It may be argued that the vector group is made more conclusive if a resultant is constructed. However, by this construction a fictive wind direction is introduced, which may lie far from a possible dominant direction, and which does not show the character of the azimuthal distribution. A method which illustrates the danger of using fictive directions has been proposed by Davidsson (1956). He constructs the directional resultant and from that the components perpendicular and parallel to the coast, "in order to get a measure of transport parallel to the coast and abrasion perpendicular to it" (op. cit., p. 155). The meaning of this construction is doubtful, and must be erroneous as expressed, because the abrasion cannot be represented by a component perpendicular to the shore, which is fictive in relation to the real waves, and nor is the balance parallel to the shore a measure of intensity in transport. If, for example, there is a locality with strong oblique winds, but only weak winds perpendicular to the shore, and the strong winds are rather balanced, the result of the construction will be a great component of "perpendicular abrasion" and no component of transport parallel to the shore.

Conclusion

Beach-drifting in the swash zone and longshore drift in the breaker zone are the most essential morphological activities with regard to transportation in the shore environment. Shore retreat and accretion are closely related to these movements. There is as yet no method by which the rate of transport can be calculated with a good approximation, and for the main part of the coasts quantitative estimations cannot be made because of insufficient wind and wave data.

Wind-frequency or wind-force distributions do not involve wind duration and fetch length, both essential variables as regards wave generation. Because different fetch lengths and durations are required for different wind velocities to generate a fully arisen sea, formulas of wave force, based on wind frequencies and shore-to-shore fetches, are not entirely correct. However, wind-force and fetch distributions seem to characterize in a qualitative way the relative distribution of potential wave force acting on the shores. The fetch distribution is of great importance in cases of very unequal fetches especially if the directions of large fetches are correlated to the extreme wind velocities and long durations.

The functional meaning of the resultant of wind work, according to Schou, is indistinct as concerns coastal development. In my opinion the effect on shoreline configuration due to the wind is most correctly represented by the whole group of wind-force vectors facing the shore. Only wind velocities generating waves which have an appreciable transport or abrasion capacity should be counted. The exponent of the wind velocity which best represents the effect on the shore is uncertain, but when comparisons are made only within a vector group, the problem of the best exponent is not so important, as long as the wind-frequency distributions of the different directions are of the same type or the frequencies giving the different magnitudes of the vectors are mainly within a limited range of velocities.

The vector group should be looked upon as a first approximation to one of several basic elements necessary to a discussion on the present shoreline morphology. Its principal advantage lies in being the most correct expression possible, when only ordinary wind statistics are available. In investigations on fossil shorelines or coastlines with a long and slow development, the possibility of changes in the wind distribution must be kept in mind.

INVESTIGATIONS ON THE SHORE MORPHOLOGY OF LAKE VÄTTERN

Wave action in profile and sediment distribution

Introduction

In this section a sample of depositional profiles will be discussed. The shore zonation has not the complexity of tidal areas. The division with regard to form and sediment distribution which can be applied is similar to the one used by Inman (1953) on beaches at La Jolla, California, in which the division is based on the particle distribution (Fig. 46). He distinguishes four types: Type I, beach foreshore sands, the best sorted; Type II a, the surf zone with coarser material than the former and more poorly sorted than the sediments inshore and offshore; Type II b, predominantly well-sorted fine sands on a relatively level shelf out to depths of approximatively 100 ft; Type III, found seaward of Type II b in areas, where the slope is steeper than about 1:15. These "slope" sediments usually have more silt and are characterized by positive skewness and poorer sorting (op. cit., p. 2).

In Lake Vättern the following zonation can be distinguished.

Zone I. The foreshore zone, that either consists of a beach, in which the step is included, or of an abrasional pavement without movable material.

Zone II. The inner breaker zone, which generally has a steeper gradient than the area offshore but a lower gradient than the step. The material is often poorly sorted.

Zone III. The outer breaker zone, which has a low gradient and in some areas forms a platform (shelf). Sand predominates, and in this zone the subaquatic bars are found. The outer part is only directly affected by wave motion in hard weather.

Zone IV. The offshore zone, which is characterized by sediments deposited from suspended load. Only in its inner part there is some direct wave influence. In some areas this zone can be distinguished by a steeper slope than that of the previous zone, but there is not always a distinct discontinuity in the profile. The boundary between zone III and IV is the limit of the nearshore area.

The foreshore

The profile of a sand and gravel beach

As a rule, the beaches rest on an abrasional pavement that can be identified by the digging of shafts or on an abrasional surface cut in cohesive deposits or solid rock. Where a beach has been formed on glacifluvial sand, it is often impossible to distinguish the beach material from the glacifluvial sediment *in situ*. The problem is the same as regards the separation of recently deposited material in areas entirely built up by the waves (barriers, beach ridges, etc.). Even on very recently formed beaches, as, for example, in artificially dammed lakes and rivers, it has been found to be almost impossible to state exactly the basis of the beaches, when it consists of sand. The abrasional forms are dealt with in the special section on abrasion. Here it can be said that even in the foreshore zone the abrasional surface has a low gradient, and the slope is continuous with the inner breaker zone.

A typical profile of a beach built up of sand and gravel is illustrated in Fig. 47. There is a distinct step of the coarsest material present, and off this step the sandy inner breaker zone is rippled (cf. Fig. 48). The arrow at point 2 marks the general upper limit of recent swash and point 1 the limit of occasional overtopping. At the time of the levelling and sampling the swash reached to point 4. The mean gradient above the step is 1:8.5, on the step 1:4, and off the step 1:10.5.

Samples were taken at the eight points marked on the profile and on the lower part of the step. The samples consist of the sedi-

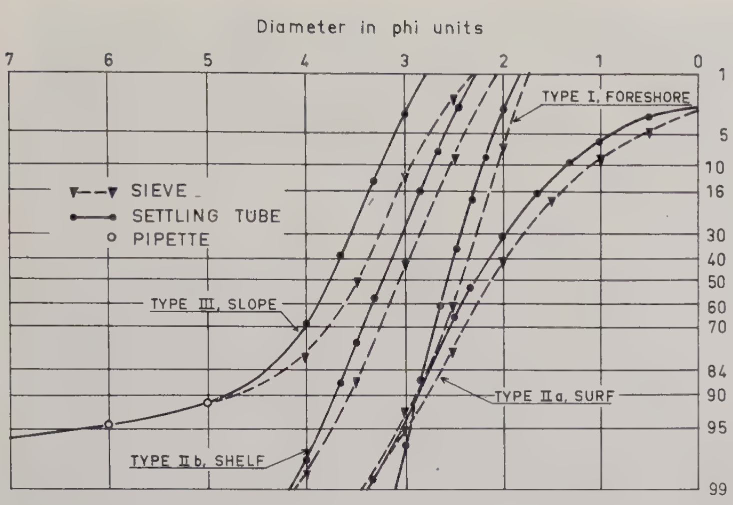


Fig. 46. Examples of characteristic grain-size distributions of the shore zones at La Jolla, California (from Inman 1953).

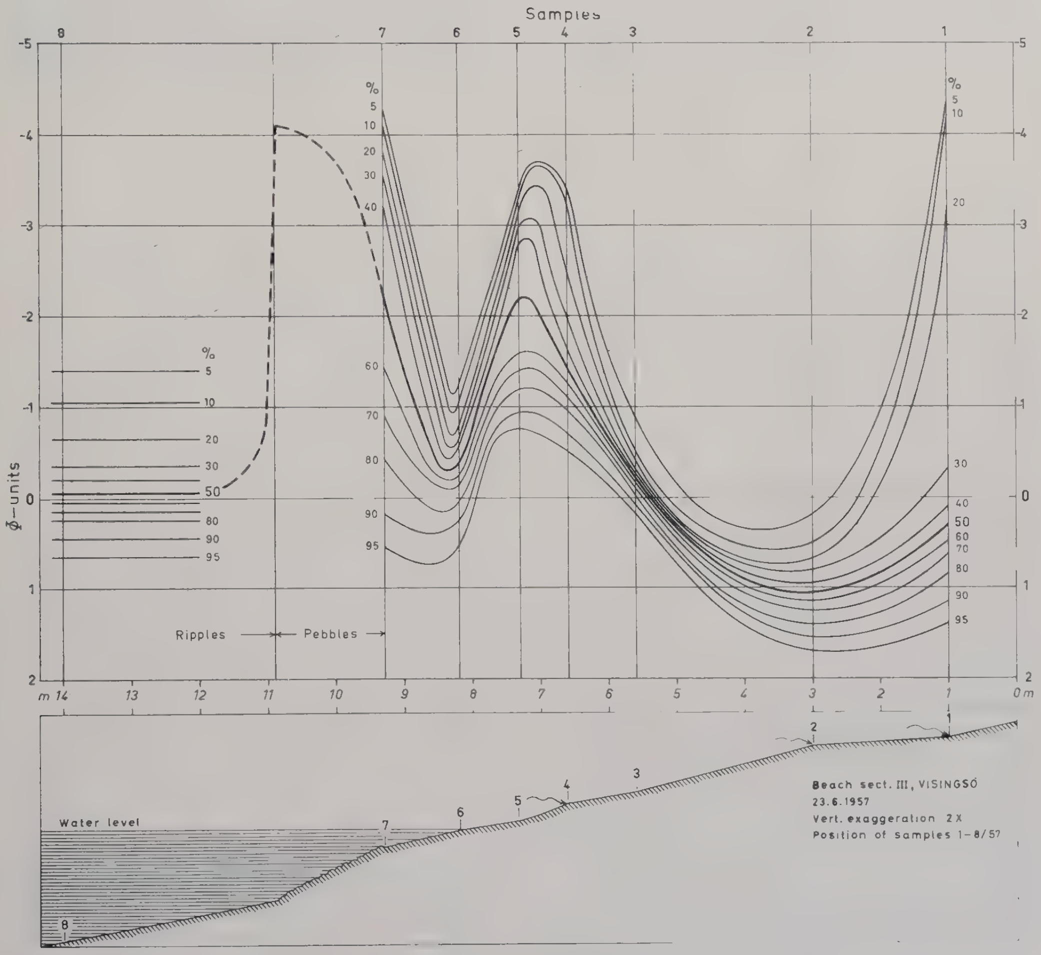


Fig. 47. Grain-size distribution in a beach profile on the northern shore of the Sandudden ness.

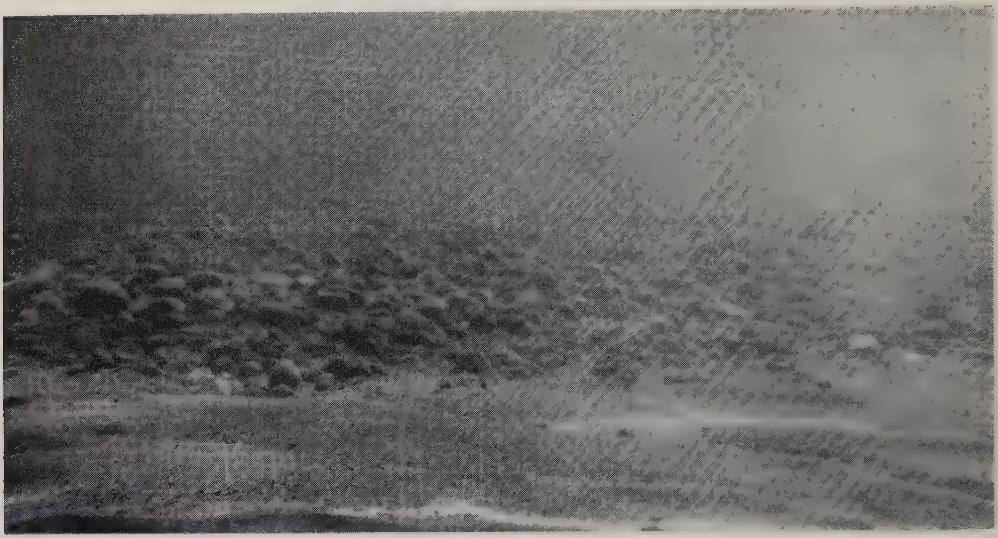


Fig. 48. The front of a step of gravel and small cobbles. Note the imbrication and the sharp limit between the sand ripples in the foreground and the coarse material of the step. Visingsö island, July 1957.

at the 28 % level. The coarse particles of these samples almost entirely consist of flat ment down to a depth of 3 cm. The diagram shows the particle distribution represented by the cumulative percentiles (the percentages coarser than the percentiles shown). The curves show the values found by the sieving analyses (cf. Table 26) at the positions of the sampling points, between which the curves are estimated. The curves are not drawn on the step, because it is impossible to take representative samples of a large enough volume in this environment. Deep grabs cannot be taken, because under the surface the distribution is not the same–it is finer. On the upper part of the step the particle size rapidly increases downslope, and the maximum sizes are found on its base. The median of 50 particles of the lower part is shown (2 cm). No particle was larger than 6 cm. However, this median is too low as regards the actual surface of the step, as the larger pebbles cover the surface (cf. Fig. 48).

From the top of the step the grain size rapidly decreases towards the mid-point of the swash in action, and the sorting becomes better, but in the upper part the median again increases. It will be noticed that the samples of points 4 and 5 have a bimodal distribution. In both samples the point of inflexion of the cumulative distributions lies at $-2.2\ \emptyset$, but at point 5 it lies at the 50 % level, and at point 4 shale pieces, which form a band along the shore at the top of the swash.

From this band up the slope the particle size gradually decreases to the top of recent swash. The sorting is best in the middle of the slope, but there is still a small proportion of coarser particles, and their importance for the curves of the distribution increases towards the ultimate swash limit. From the figure it can be seen that they constitute a distinct bimodality in the area of overtopping, where flaky particles larger than about 2 cm are found. If the shale particles were not present, a distribution with decreasing sizes and good sorting would be found.

The shale pieces are whirled up by vertical turbulent motion on the step, and are then carried by the uprush. When they settle at the top of the swash, they will not be eroded by the weak, less turbulent backwash because of their flatness. In areas where flat pebbles and sandstone exist, scattered pebbles or a band of pebbles will regularly be found on the top of beaches of quartz and felspar sands (cf. Fig. 49).

Sorting with regard to shape on shingle beaches

The relations between shape, kind of rock and distribution on the foreshore have been examined on some shingle beaches. The shape was

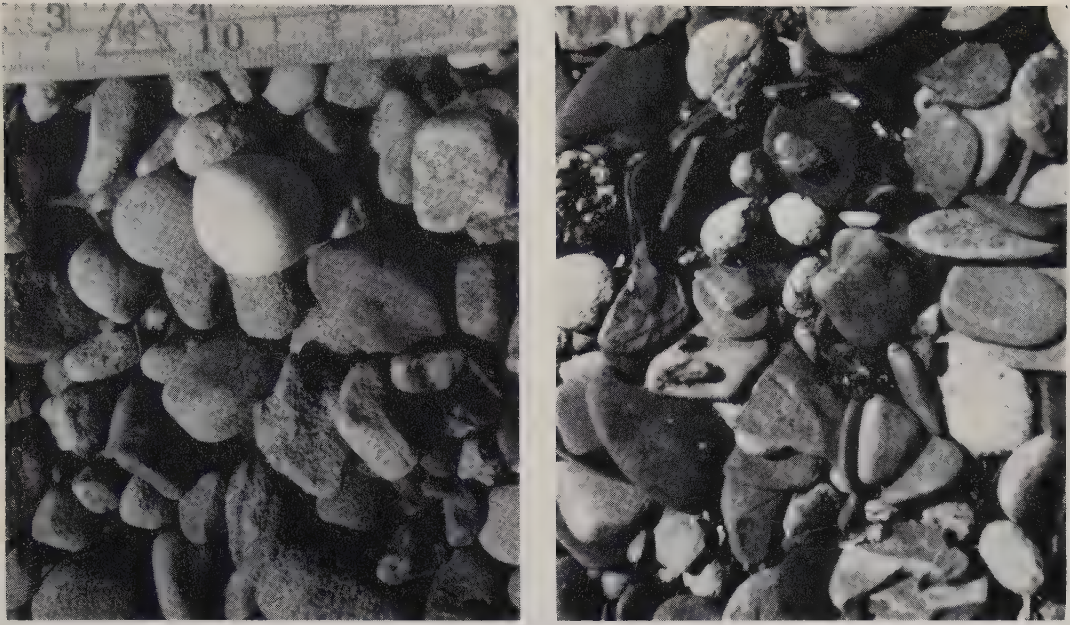


Fig. 49. Material from the same profile of a mixed gravel-sand beach. The left-hand photograph shows spheroid and ellipsoid particles at the shoreline, the right-hand photograph flat shale particles at the top of the swash zone. Centimetre scale. Visingsö island, July 1957.

represented by the long axis (*a*), the semi-axis (*b*), and the short axis (*c*), perpendicular to the b-axis. The median and the quartiles of the ratio *a*/*b* (length/width) and *b*/*c* (width/thickness) were calculated.

Different kinds of rock are originally split in specific forms, and where, as on Visingsö island, the cobbles of the moraine consist of Archaean crystalline (mainly granite) and sedimentary rocks, extremes of shapes will be represented. Shale, pelite and sandstone are originally split in flat plates. The sandstone plates are generally thicker than the other ones. The thin plates are more likely to move in a sliding motion than to roll, and by attrition they get a disc shape. Arkoses and even sandstones often split in parallelepipedic forms. These can roll over the semi-axis and obtain an ellipsoid form. The limestone (of the Visingsö Series) has a tendency to split rather cubically, and the cubic pieces are worn to spheroid or egg-shaped forms. The granite pieces of the moraine usually have no predominant flatness, and they get an ellipsoid or spheroid form.

If the form is of predominant importance to the sorting with regard to the position on the foreshore, a co-variation with regard to the kind of rock can also be expected, but rock pieces which have a form deviating from the regular of their kind can be expected to be found together with rocks for which this form is the normal. The relations will here be exemplified by measurements from the beach south of Sandudden on Visingsö island. On this beach the following rock types were distinguished: crystalline rocks, mainly granite (C), arkoses (A), sandstones (S), and pelite and shale (P + Sh). Samples were taken at the principal updrift end (the southern end), in the middle of the beach, and at the downdrift end. At each locality separate samples were taken on the step, on the mid-slope just above the waterline, and on the upper part of the foreshore slope. The mid-slope of the downdrift end was covered with gravel and sand.

The distribution with regard to the kind of rock is seen from the first columns of Table 13. The crystalline rock pieces predominate on the step and on the mid-slope, especially in the updrift and middle part of the beach, and there are few or no pelite and shale pebbles and cobbles. On the upper slope the relation is the reverse, but in the updrift area there are

Table 13. Morphometry of foreshore shingle on a beach of Visingsö island.

The updrift end

	Rock		a/b			b/c		
	Spec.	%	Q_1	Md	Q_3	Q_1	Md	Q_3
Step	C	57	1.23	1.32	1.48	1.30	1.38	1.58
	A	23	1.12	1.25	1.41	1.18	1.85	2.33
	S	20	1.08	1.35	1.48	1.13	1.30	1.65
	P+Sh	0	—	—	—	—	—	—
	—	100	1.17	1.31	1.46	1.24	1.34	1.77
Mid-slope	C	41	1.18	1.33	1.50	1.18	1.24	1.38
	A	32	1.13	1.30	1.54	1.09	1.30	1.45
	S	25	1.14	1.19	1.46	1.34	1.52	1.68
	P+Sh	2	—	—	—	—	—	—
	—	100	1.15	1.28	1.50	1.19	1.33	1.48
Upper slope	C	22	1.28	1.36	1.45	1.13	1.19	1.43
	A	20	1.17	1.26	1.55	1.08	1.35	2.13
	S	46	1.10	1.33	1.53	1.30	1.48	1.70
	P+Sh	12	1.29	1.38	1.47	2.04	4.07	7.42
	—	100	1.18	1.33	1.51	1.31	1.70	2.41

The middle part

Rock		a/b		
Spec.	%	Q_1	Md	Q
C	54	1.20	1.33	1.5
A	26	1.15	1.28	1.4
S	17	1.15	1.33	1.4
P+Sh	3	—	—	—
—	100	1.18	1.32	1.4
C	55	1.10	1.28	1.4
A	15	1.15	1.25	1.4
S	21	1.16	1.33	1.4
P+Sh	9	—	—	—
—	100	1.12	1.29	1.4
C	13	1.11	1.18	1.2
A	18	1.22	1.35	1.5
S	26	1.15	1.30	1.4
P+Sh	43	1.15	1.24	1.3
—	100	1.16	1.27	1.4

still many crystalline pebbles, and sandstone predominates.

The length/width ratio has a small variance. The medians of the updrift end and the middle of the beach lie within a narrow range. In the downdrift end they are somewhat higher.

The distribution of the width/thickness ratio is of the greatest interest. At each locality the difference between step and mid-slope is small, but on the upper slope the flatness is significantly greater. The flatness on the upper slope of the updrift end is the same as that on the step of the downdrift end, and it may be noted that these two samples have about the same composition with regard to the rock species.

An examination of the flatness distribution with respect to the individual rock species shows that on the step of the downdrift end the arkoses and sandstones and crystalline pebbles have about the same flatness as on the two other localities. The pelite and shale pebbles are much rounder than those found on the upper part of the foreshore. It is notable that the few crystalline pebbles of the upper foreshore are unusually flat. At the middle part of the beach the few crystalline pebbles of the upper slope are flatter than the ones in lower positions. The ratio of the arkoses also increases up the slope. The number of shales and pelites in the lower part is too small to be counted (the median at the mid-slope is 4.7), but the pieces of the upper part are very flat. At the updrift end there is no clear tendency in the sorting perpendicular to the shore. The shales and pelites of the upper slope are poorly sorted.

With regard to lateral sorting the following general trends can be found from the table. In the downdrift direction there is an increase of flatness on the step, owing to the sandstone, shale and pelite pieces. On the mid-slope there is a general increase to the middle of the beach, and this increase is pronounced along the upper slope. There the sandstones and shales are flatter in the middle part than at the downdrift end. This can be interpreted as an effect of less transportability of the flattest material by longshore drift.

Because the percentual composition of the rock species varies considerably from one area to another, it is difficult to state general quantitative relationships between shape, dimension and position on the beach. However, from the

b/c		
Q₁	Md	Q₃
1.28	1.45	1.73
1.50	1.90	2.90
1.53	1.85	2.40
—	—	—
1.38	1.64	2.16
1.24	1.42	1.72
1.58	2.35	3.53
1.72	2.15	3.22
—	—	—
1.41	1.74	2.36
1.42	1.85	2.09
2.48	3.50	4.46
3.36	4.05	5.05
4.72	5.70	6.44
3.53	4.37	5.16

The downdrift end

Rock		a/b			b/c		
Spec.	%	Q₁	Md	Q₃	Q₁	Md	Q₃
C	27	1.25	1.44	1.60	1.15	1.27	1.47
A	19	1.23	1.38	1.50	1.37	1.82	1.98
S	32	1.23	1.34	1.63	1.29	1.93	2.18
P+Sh	12	1.16	1.42	1.78	1.76	2.25	3.77
—	100	1.23	1.39	1.61	1.31	1.70	2.07
Gravel and sand							
C	6	1.25	1.40	1.80	2.06	2.30	3.30
A	26	1.30	1.41	1.50	2.34	2.83	4.18
S	26	1.32	1.40	1.51	2.22	3.23	4.69
P+Sh	42	1.25	1.45	1.69	3.54	4.36	5.54
—	100	1.28	1.42	1.60	2.79	3.54	4.83

studies on Visingsö island, where comparatively large quantities of shale and sandstone are found, some general trends can be stated. The flat pebbles and cobbles are regularly found on the upper part of the slope, and the more ellipsoid and spherical on the lower part. The initial forms, depending on the kind of rock, are accentuated by the attrition. The flat particles slide up and down, while the cubic or parallelepipedic ones have a greater tendency to roll. The rolling particles are more effectively helped in a downslope motion by the gradient of the foreshore than the sliding particles. On the other hand, the flat particles have a larger lifting surface in relation to weight, and are more easily carried upslope by the turbulent uprush. Therefore, on the upper slope the flat particles are larger (with regard to the length of the smallest projection) than the ellipsoid particles.

Particle orientation on shingle beaches

On shingle beaches distinctly preferred orientations of the particles can be found. There are two types of orientations; one is dependent on the way the particles are transported, and the other on the reaction in the swash of particles not transported.

It is well known that, with regard to inclination, shingle on the middle and upper foreshore generally tends to be imbricated towards land like an inverted roof slating. The material is predominantly affected by the uprush but not transported. This orientation has been found to be comparable with that in rivers (cf. Guilcher 1958, Fig. 5 C). However, the reverse inclination with the axis inclined off the shore is found on the step (cf. Fig. 48) as a result of the downwash and the mainly lifting action at the start of the uprush.

The horizontal orientation will be illustrated from a small bay in Lake Vänern, because this locality illustrates the general conditions extremely well (Fig. 50). The measurements were made by Miss G. Wennlund on my instructions. They were made in the area from the shoreline to 1.5 m above it. The measurement of the length axis varied between 8 and 15 cm. The bay is predominantly exposed to waves from south to south-west. The western part is thereby somewhat more sheltered.

In all the three diagrams there is a north-southerly maximum. This is caused by the

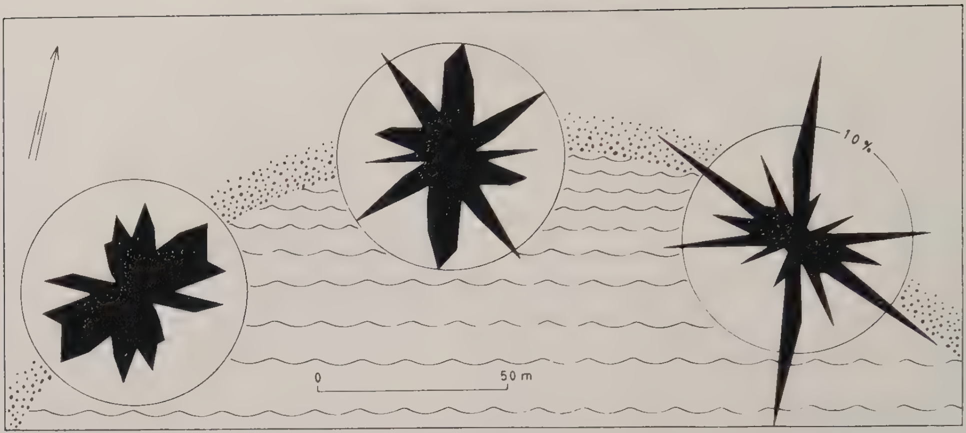


Fig. 50. Long-axis orientation in the horizontal of small cobbles in a bay of Brommö huvud, south-western Lake Vänern. Measurement by G. Wennlund 1958.

prevalent waves, and the particles included in this maximum are the ones inclined towards land and not transported during the orientation. There are two secondary maxima with an angle of about 45° to the shoreline. These have been generated in the same way as the first one, but by oblique swash from south-west and south-east. Finally there is a maximum of particles orientated along the shoreline. This maximum is made up of particles, which have rolled on the slope, and the main part of them is found at the shoreline. This rolling movement has been observed in action on several beaches. Their orientation is thus less dependent on the swash direction than on the slope, which determines the rolling motion during the downwash. Finally it should be remarked that an alongshore orientation caused by the uprush can also sometimes be found on the top of the foreshore.

Notes on shingle distribution

In Lake Vättern the coarsest, well-rounded, crystalline pebbles and cobbles are found in "pocket" beaches along Mount Omberg and in the area north of the Svedudden promontory. By "pocket" beaches is meant small beaches, which are isolated by rock projecting into the water off the inner breaker zone. Thus material cannot enter the beaches or be lost by longshore drift. During the submergence the material has been kept on the shore and the rate of supply from the cliff has been slow. The well-polished surfaces and the ellipsoid form are illustrated by a sample from the Omberg shore (Fig. 51). In the pocket beaches north of the Svedudden promontory the well-rounded material at the shoreline has a size of 8 to 15 cm (cf. Fig. 52).

In freely exposed areas, where the water depth is at least 1 m in the inner breaker zone (as in the Svedudden and Bankeryd areas and the Jönköping and Huskvarna Bays), sand and shingle are thrown up to an average maximum height of 1.2 m above the mean water level (cf. Granlund 1934 a, p. 10). Under favourable conditions the wave run-up can be far higher. However, in sheltered areas (such as, for example, the Motala bay and the northern archipelago) and in areas, where the nearshore zone is shallow (the area west of the Flisen shoal, the northern and western shores of Visingsö and the eastern shore of Stora Röknen island), the average maximum height of the foreshore deposits is not more than 0.4 to 0.6 m.

The inner breaker zone

The bottom of this zone is either an abrasional suface or there is material deposited by the wave action. The deposits either consist of rippled sand, poorly sorted gravel and sand, or cobbles and boulders.

Rippled sand is mainly found in sheltered areas or in areas where there is a large accretion. The sediment type of coarse sand and gravel is most common. The pebbles are transported with the finer material (Fig. 53),

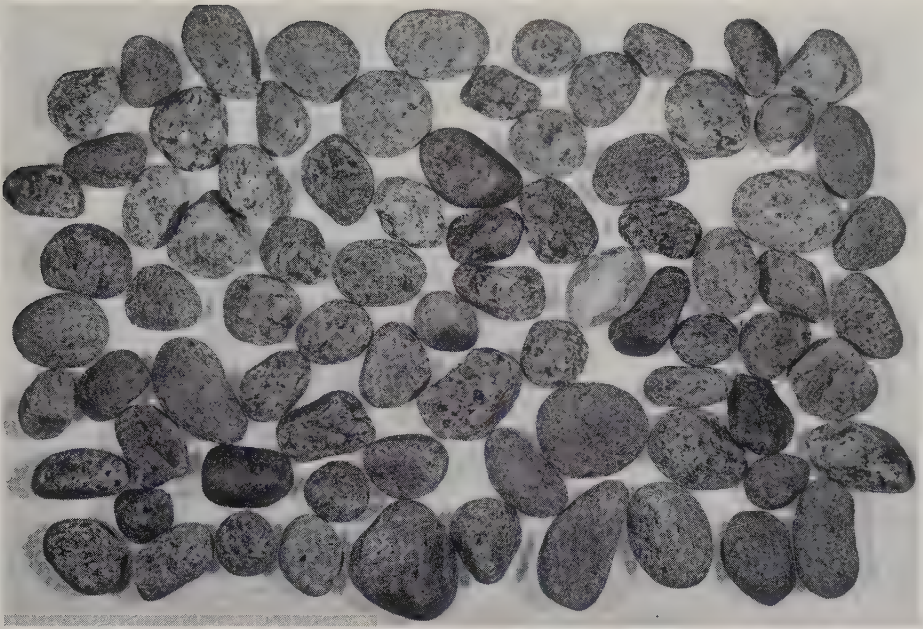


Fig. 51. Granite pebbles from a pocket beach of Mount Omberg. Ellipsoid and pear-shaped forms predominate. 25 % of natural size.

but in updrift areas of longshore drift there will be an enrichment of the coarser particles until a covering pavement is formed (Fig. 54). The mean gradient of 20 profiles with this sediment pattern was found to be 1:8.1.

When the material consists of cobbles and boulders, it may be difficult to judge if it is a true residual material or if it has been transported by the waves. The material, which forms an abrasion pavement over vast shallow areas

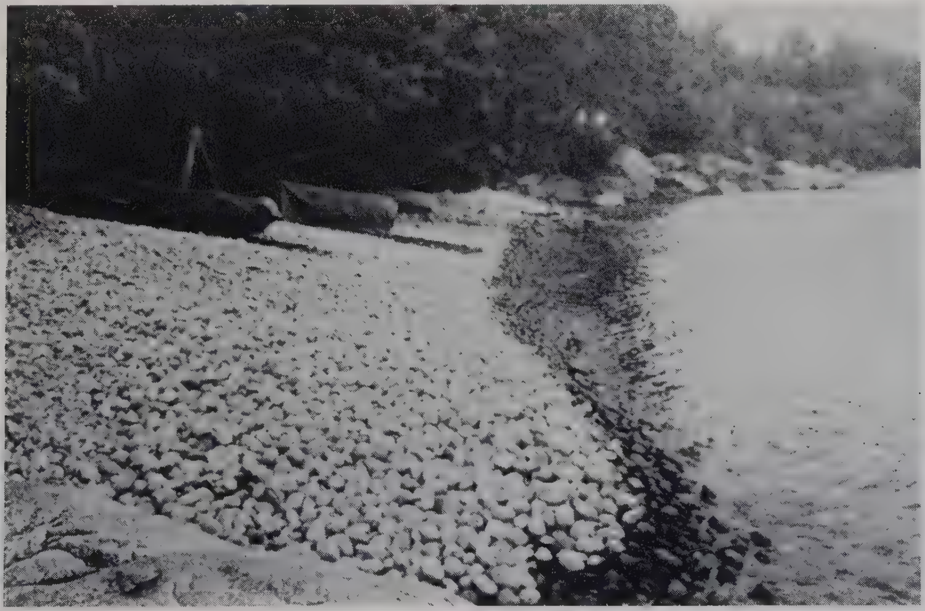


Fig. 52. A pocket beach with granite cobbles in Brandstorp on the western mainland shore. Aug. 1962.

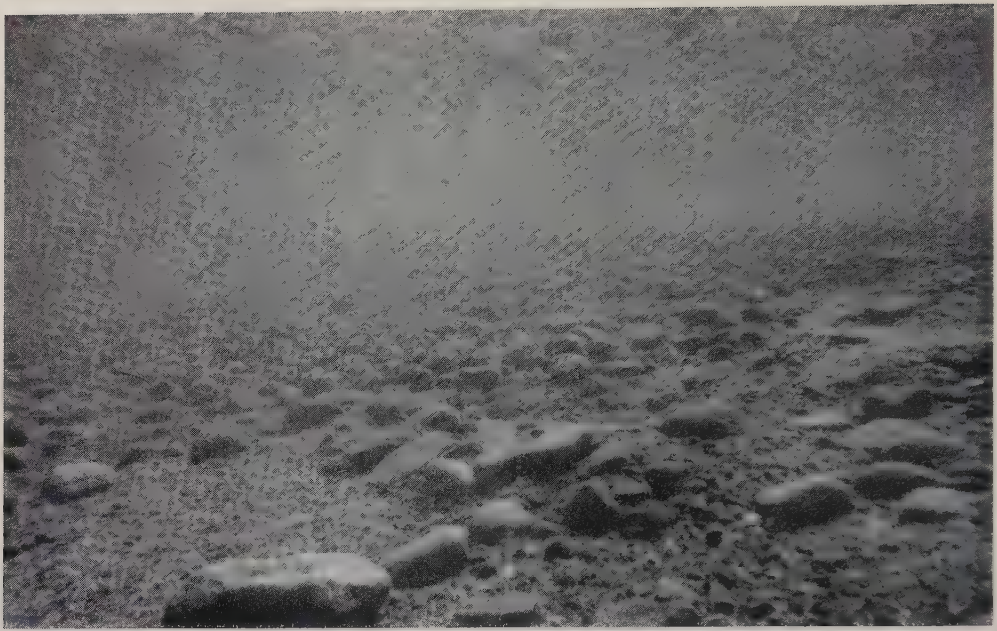


Fig. 53. Transport surface of pebbles and small cobbles on gravel and coarse sand at the Sandudden ness. View contrary to direction of drift. Depth in the foreground 0.7 m. July 1957.

north and west of Visingsö island, has only an average size of 10 to 15 cm. There the gradient is low, and the material rather flat. On the narrow platform along Mount Omberg a distribution has been examined, which shows an increase of size from about 1 dm at a depth of 0.5 m to 5 dm at a depth of 2 or 3 m. There the material has a round shape, although partly

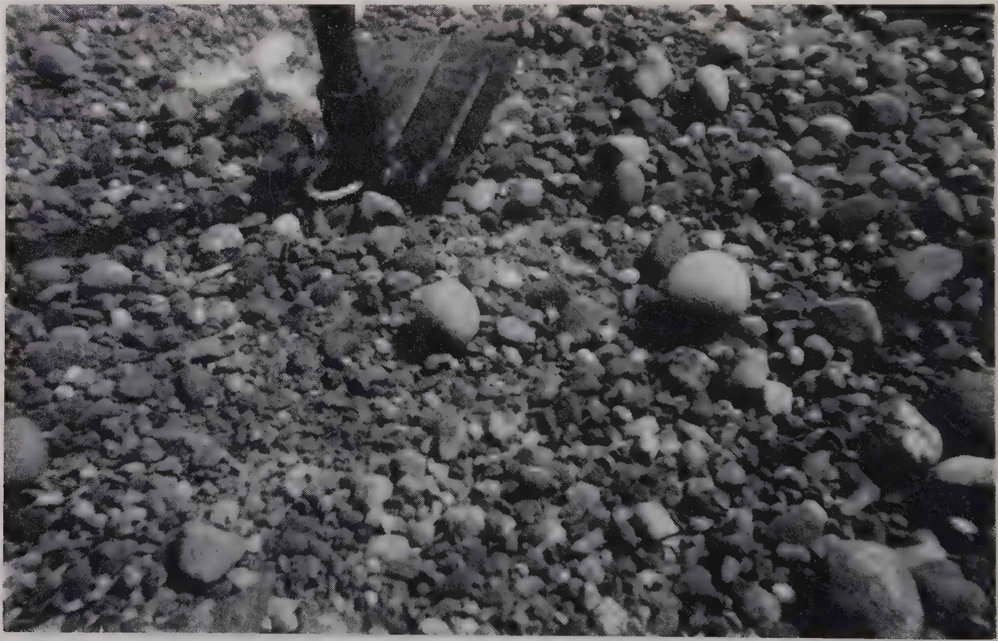


Fig. 54. Close-up view of a transport surface of pebbles and gravel at the Sandudden ness. Depth 0.5 m. July 1957.

angular. This distribution must be regarded as having been actively influenced by the wave action.

In northern Vättern several well-exposed shores with a steep inner breaker zone can be found. They have been visually examined, but few measurements have been made. The gradient is of the order of 1:4 to 1:8, and there are frequently boulders of the size of 2 to 3 dm. They have been washed out of moraine but have probably only taken part in a limited transport.

From observations of recent rock-fills on the Svedudden promontory it is certain that half-metre blocks can be drawn down on steep slopes. For a road bank along the south-western shore of the lake, Hedar (1960 *b*) has found by experiments that a block weight of 300–500 kg is necessary for a gradient of 1:2.7, when the depth does not exceed 1.8 m. The same stability could be obtained for depths not exceeding 3 m, if the gradient was lowered to 1:3.5.

Thus the waves can be expected to have a destructive effect even on very coarse material on steep slopes, but when the material has been drawn down and the profile is lowered, the movements will cease. From many notes about the maximum size of well-rounded material it can be estimated that the maximum size effectively transported and abraded is about 2 to 3 dm. This size agrees rather well with the size which could be expected from the experiments by Ippen and Eagleson (cf. Fig. 41).

The outer breaker zone and the offshore zone

General

The two outer zones of the profile will be discussed together, because the relations between them are of importance. There are different opinions about the processes, which determine the morphology and the sediment distribution, and thus these zones are of a general interest. Unfortunately, the wave conditions of Lake Vättern can only be discussed rather qualitatively.

With regard to the local problems, the present conditions are most important for the interpretation of the sediment distribution of the submerged areas and the determination of the lowest shore levels. These problems, on the other hand, affect the possibility of investigating the relations between existing morphology, sediment distribution and present shore processes.

In the whole area south of the northern archipelago there is always the risk that the sediment distribution in the deeper parts of the outer breaker zone and in the offshore zone may reflect a lower shore level than the present, and in the archipelago the conditions with regard to wave exposure and sediments are not representative of the rest of the lake. The effects of the unequal uplift and the distribution of the Quaternary deposits thus restrict the useful areas. Much preliminary surveying has been done to find a suitable area for the study of recent wave activity and sedimentation. It was found that the Jönköping bay was the best area for these studies. This may be thought paradoxical, because it is the part of the lake with the maximum submergence expected (cf. Fig. 7), but it is the only area where large amounts of material have certainly been continuously deposited, and the distribution of this material depends on the recent wave action.

In the following sections the general morphological character of the outer breaker zone and the offshore zone in Lake Vättern is first described, then some fundamental characteristics of the particle distributions are discussed, and in the two last sections the Jönköping bay and its sediments are described and discussed.

Morphological characteristics

The morphology most common is not depositional but abrasional, and consists either of an abrasion pavement or an abrasion surface, cut in sedimentary rock or cohesive Quaternary deposits. Large areas of nearshore sand have been found in two different types of position. One is the well-developed bays: the secondary bays of the Motala bay, the bays south of Karlsborg and the Jönköping bay. The other is the shallow platforms: off Gränna, the Sandudden spit on Visingsö island, and the area south of the Svedudden promontory on the western shore.

In the bays subaquatic bars are regularly found. There are always at least two, generally three, parallel bars. Three profiles from the Motala bay and one from the Jönköping bay are shown in Fig. 55. It may be noted that the bars of the Motala area lie in much shallower

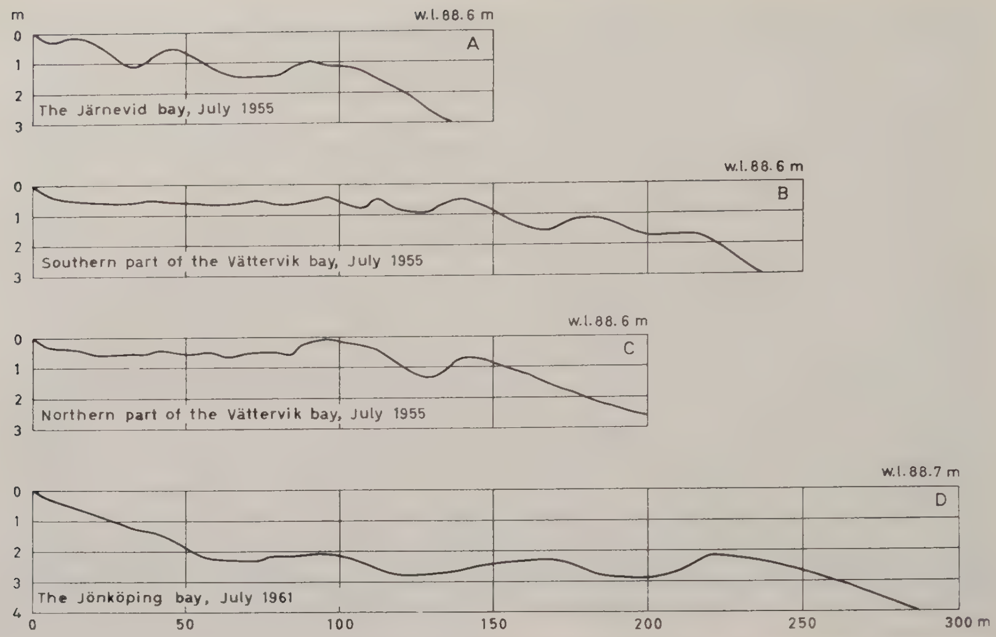


Fig. 55. Profiles of subaquatic bars. The profiles *A—C* from the Motala bay section levelled by L.-O. Johansson.

water and that their lengths are shorter, but their heights are almost the same or somewhat higher. The inner bar of profile C has a height and a form which make it questionable, whether it is not really a swash bar. Similar forms are found in the very shallow Varamo bay north-west of Motala.

In shoreline inundations (which can be classified as open bays) west of the Flisen shoal, south of Hjo, and along the western shore of Visingsö island, there is a tendency for bars to form, even though there is not enough sand present. The sand moved over the abrasion pavement of the bottom is accumulated in two or three separate strips, parallel to the shore. Their height is at most about 10 cm, and is generally only a few centimetres. These embryonic bars are best developed in the downdrift part of the open bays, when an oblique steady wind has blown for some days.

On the previously mentioned platforms without open bays there are no subaquatic bars, although much sand is present. The bottom is very even, but low sand banks with a diffuse horizontal configuration have been sounded. Because the bottom is rather flat (cf. Fig. 10 and Fig. 71), the predominant waves moving along the lake are little refracted, except close to the shore, and thus most of the transport can be judged to be along the shores, conditions which seem to prevent the formation of bars.

The offshore zone has no general specific morphological character. In areas where there is a nearshore wave-built platform, this zone is marked by an increase in slope and a decrease in grain size. However, large areas of the central and southern lake sections consist of abrasion pavements formed during lower water stages. The bottom slope is continuous, or there are large steps, which can be attributed to the lower shore levels (cf. Fig. 145). In the northern section bottom sediments of Late-Glacial origin are frequently found.

If all the waves affecting a shore have a unimodal distribution with a limited variance, there will be a specific range of waves, which has a predominating effect on the morphology. A unimodal distribution does not exclude large seasonal variations (large steep waves in winter, small low waves in summer). With a distribution composed of waves generated by ordinary weather situations and occasional, extremely hard winds, lasting effects can be expected from the latter.

Waves can be regarded as having a destructive or constructive effect, depending on their steepness, but it can be assumed that the wave distribution with regard to breaking depth determines the level of the nearshore morphology. If the energy forwarded to a shore is distributed over a wide range of wave sizes, the correlation is probably poor, but if there is a well-defined energy maximum, this can be expected to be of predominant importance for the position of the nearshore forms.

The total wave energy (E_L) per unit breadth of wave crest per wave length is

$$E_L = 0.64 \ w \ H^2T^2, \tag{34}$$

where w is the weight of the water per unit volume (cf. O'Brien 1942). Half of this energy is kinetic and half is potential, and the total energy of the wave train moves at half the velocity of the wave form. Thus, per unit of breadth per time unit, the following energy (E) is forwarded,

$$E = 0.64 \ w \ H^2T^2 \ 0.5C/L = 0.32 \ w \ H^2T. \tag{35}$$

If the wave-height distribution and the corresponding wave periodicity are known, the total energy distribution, related to the wave height, can thus be calculated, and with known steepnesses it can also be related to the breaking depth.

There are no data detailed enough to permit a true quantitative analysis of these relations in Lake Vättern, but a merely qualitative attempt will be discussed. The frequency distributions of wind velocities in the intervals NW.–NE. and SW.–SE. of Visingsö, and NW.–NE. of Jönköping are known (cf. Fig. 17). It is assumed that the northerly intervals have an average effective fetch of 30 km and the southerly interval an average effective fetch of 13 km, and further that the winds of the velocity intervals 4.5–7.5 m/sec to 19.5–22.5 m/sec raise waves corresponding to the velocities 6 m/sec to 21 m/sec and the minimum duration of the fetches. By multiplying the velocity frequencies of each interval by H^2T, a quantity which represents the relative energy is obtained (H is the significant wave height). In Fig. 56 these quantities are related to the breaking depths of the significant waves. The method is, of course, crude, because the durations are not known, but, because the fetches are limited, the minimum durations required are short.

The graphs of Fig. 56 show that in the northerly intervals most of the energy is connected with breaking depths of about 1 to 2 m (wind velocity interval 6–12 m/sec). The depths are somewhat larger for Visingsö. For breaking depths greater than about 2.2 m the energies fall off rapidly, but for a depth of 3 m the energy is still about a quarter of the maximum. The southerly interval of Visingsö has a well-defined energy maximum, dependent on the high frequency of 9 m/sec winds. The small depth (0.9–1.0 m) is due to the short effective fetch.

The crests of the subaquatic bars of the Jönköping bay lie at depths between 2.2 and 2.4 m, and the troughs between 2.35 and 2.9 m. With an onshore wind (cf. Fig. 32) and high water stage these depths will increase by 0.2 to 0.3 m. That is to say, the position of the bars corresponds to the part of the graph that is just above its energy maximum. On Visingsö island there are no bars, but the area down to a depth of 2 m is broad, and the main part of the moving sand east of Sandudden is found down to about three times this depth. With regard to the fetch, the SW.–SE. graph can be applied to the Motala bay. The bars of this area are found in positions around the 1 m level, which corresponds well with the graph.

Sediment-particle distributions

Statistically, the particle distributions are generally characterized by the median grain size, a measure of the dispersion (sorting) and a measure of the skewness. Because the sediment deposition rather depends on the terminal fall velocity than on the geometric size, measures based on the logarithm of the size are preferable (cf. Krumbein 1936). Even though it would be more consistent to use settling velocity as such (cf. Ippen and Eagleson 1955), this is generally not done because of the general use of the sieving method as regards coarse material.

It has been argued that a measure of sorting, based on percentiles, is only a statistical description, not truly correlated to sorting in a genetical sense (Spencer 1963), and that the sorting of clastic sediments is only recognized by the degree of truncation of the distribution.

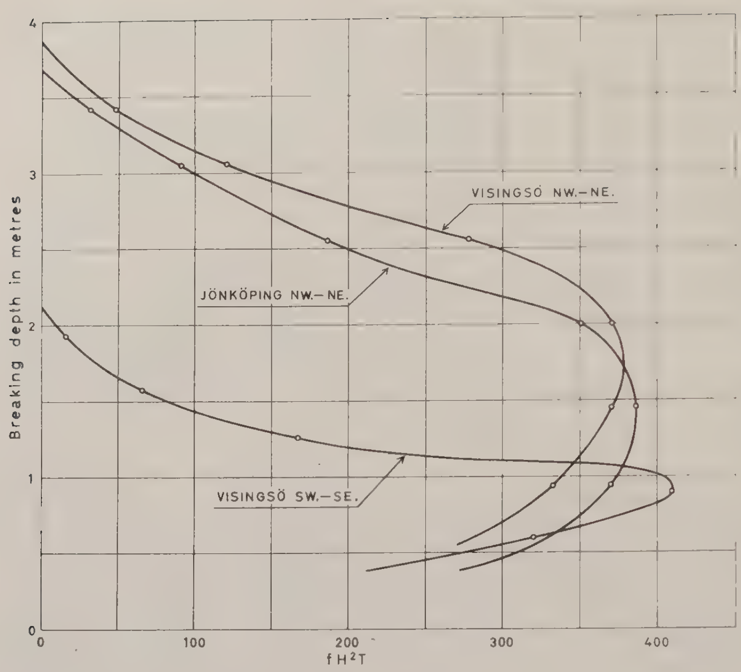


Fig. 56. Wave-energy distributions in relation to breaking depth for waves of specific azimuth intervals.

The truncation effect is, of course, not a new idea. The distribution of erosion pavements has a lower truncation, and the material withdrawn a corresponding upper truncation. Spencer's studies are mainly based on sediment analyses from the Barataria bay of the Mississippi delta, published by Krumbein and Aberdeen (1937), but he also refers to published analyses from other types of environments (not cited). From the Barataria material he finds that all bimodal distributions can be separated into two log-normal populations: a "sand" population (Md 1.5 to 4 0, σ 0.4 to 1.0) and a "clay" population (*Md* 7.0 to 9.0 0, σ 2.0 to 3.0). From other environments a "gravel" population is added (*Md* – 3.5 – 2 0, σ 0.7 to 1.5). These three log-normal populations are said to be the fundamental populations, which constitute all sediment distributions which form by mixture bimodal or polymodal distributions. These can always be separated into the fundamental distributions.

When bimodality is found in an analysis, the normal reaction is to imagine that two strata have been mixed in the sample. If the same bimodality with respect to the grain size of the inflexion point is repeatedly found, it can be considered that the sediment has been deposited under two specific, different conditions. This can, for example, be found in varved clay (Sauramo 1924). It can be questioned if the Barataria bay samples do not show an alternation between predominant bed load and suspended load, each reflecting specific conditions of this area, but it is difficult to understand why another gap between the two populations could not exist in an area of different conditions. In fluvial morphology the most important point of division lies at the lower limit of possible bed load, characterized by the grain size 0.15–0.2 mm, which may be called the critical grain size. In the nearshore zone, affected by waves, suspended material is whirled up, but it is not necessarily transported out of the environment, but may settle when the wave motion ceases. As in fluvial conditions, the finest material can be expected to be sorted out in the updrift area of longshore drift and be transported to the downdrift area, but there the energy level may be as high or higher than in the updrift area. Then we have the possible combination of large amounts of rather fine material and high-energy conditions.

In Fig. 57 nearshore bottom samples from Lake Vättern have been arranged in relation to the depth below the present water level and

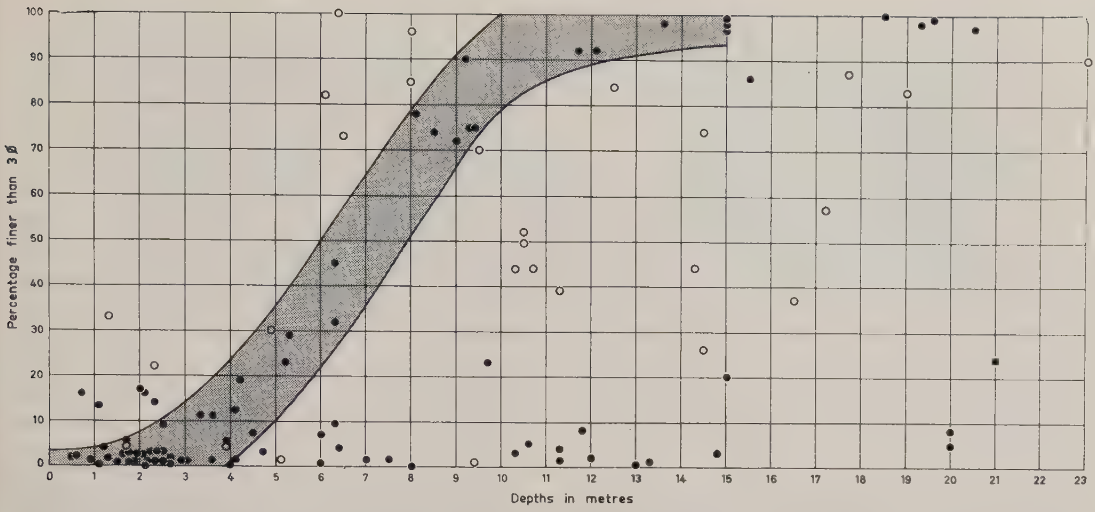


Fig. 57. Relation between depth and percentage finer than 3 Ø in samples from southern Lake Vättern. Unimodal distributions, black dots; bimodal distributions, open circles. The shaded belt marks the normal trend of variation with depth of sediments with a unimodal distribution. The sediments to the left of the belt are found in areas with a large supply of fine particles through shore retreat. The sediments on the right-hand side were deposited under special conditions, and the coarse ones found in large depths certainly at time of a lower water level.

the percentage of particles with a diameter smaller than 3 Ø (0.125 mm). These samples may represent the present conditions, or they may have a fossil character, having been deposited during periods of a lower water level.

The phi unit chosen is close below the critical grain size, and particles finer than this can be expected to represent suspended load. In the diagram unimodal and bimodal distributions are distinguished, and there is an obvious relation between the percentage of fine material and modality. In the interval with 0–30 % finer than 3 Ø five samples of 70 have a bimodal distribution, in the interval with 30–70 % 10 samples of 12, and in the interval with 70–100 % 11 samples of 28. Here we shall only note that unimodal distributions with about 30–70 % finer than 0.125 mm are scarce, and the reasons for this will be discussed in the section on the sediments of the Jönköping bay. The samples from the greatest depths with extremely little fine material are fossil sediments, and are discussed in the chapter on the bottom morphology.

If only the unimodal distributions are taken into consideration and the coarse sediments not correlated to the present water level are excluded, a tendency to a decrease in grain size can be found in the interval between 6 and 8 m, where the amount of particles finer than 0.125 mm increases from 30 to 70 %.

The sediments of the Jönköping bay

The shores

The deep trench of the southern Vättern basin is directed towards the Jönköping bay (Pl. 1), and as previously mentioned, the depth to bedrock on the shore in the eastern part of the bay is 183 m. The shore consists of two tongues, separated by the outlet from Lake Munksjön. To what extent the deeper parts of the sediments are Glacial is unknown.

The area west of the outlet was built up during the deglaciation and the period of falling levels of the ice lakes. In excavations for new blocks of flats, built in the last few years, the stratification of the Baltic Ice Lake shore has been exposed at levels between 100 to 90 m above sea-level.

The ground east of the outlet has, on the contrary, mainly been built up during the present period of submergence, and has the character of a bay mouth barrier that has embayed the lagoons of Lake Rocksjön and Lake Munksjön. The ground lies nowhere more than about 2.5 m above the lake level, but as most of the higher parts were built over already in the

17th century, the natural surface is generally lower than the apparent. The formation of this bar by gradual deposition of beach material is well illustrated by the studies of peat layers superimposed by sand, carried out by Granlund 1934 *a*).

The shore line of the bay runs in an even curve with a rather straight central part perpendicular to the axis of Lake Vättern. The present situation of the shore line is, however, of recent origin and only in part natural.

A town map from 1625, when the area east of the outlet was still unsettled, shows the shore of the eastern bar lying about 50 m further south than the shore west of the outlet (cf. Björkman 1918, Fig. 27).

In a discussion on the abrasion at Rosenlund east of the bay and the drift of eroded material, Allvin (1839, p. 37) says: "That the larger part is brought into the Jönköping bay is clear by the bank of sand that has been accumulated outside the entrance to the town (that is, the outlet from Lake Munksjön), thus making it impossible for larger ships to call without unloading some part of the cargo, and which was the main reason for the building of the outer harbour. The mentioned bay seems also to be more and more filled up, and the houses, which for lack of space were built out not inconsiderably into the lake, now not only stand on dry land, but small beaches have also started to grow in front of them." The harbour entrance mentioned was built in 1833–36.

A small-scale map by Weiler (1936, Fig. 5) shows an advance since the beginning of the 18th century of about 100 m immediately east of the outlet, but it gradually decreases eastwards to about 50 m. Weiler does not say to what extent this advance is artificial, but instead mentions that sand is permanently quarried from the eastern part of the beach (1936, p. 27). The balance between quarrying and refilling can hardly be accurately estimated, but the most influential man-made interference during the last 300 years was when a railway bank was constructed along the shore in the 1860s, and a stone revetment was built in front of it, covering the middle third of the bay. West of the harbour most of the beach is covered by rock-fill in order to shelter the Jönköping railway station from ice pressure and spray.

Thus a natural beach now only exists in the eastern third of the bay, where it is 10 to 15 m wide. In the active swash zone the beach material mainly consists of gravel. Further up sand predominates, but there are also scattered pebbles, and at the top of the beach there is some blown sand. No berms are found.

Considering longshore transport within the bay, the following morphological characteristics should be noted. The position of the outlet from Lake Munksjön seems to have been fixed for a long time. There are no traces of a tendency to systematic shifting, leading to an overlapping bar. Some natural shore advance has been reported (Allvin 1839), but most of the change in the position of the shore line is artificial. Since the first outer harbour was built in 1833–36 and the revetment in the middle part a hundred years ago, no beaches have developed in front of these constructions. The lack of accumulations at the harbour indicates a balance in longshore transportation in agreement with the fixed position of the natural outlet. The absence of a marked advance of the shore indicates that the supply of material suitable for beach formation has not been sufficient enough to produce such an advance under the conditions of submergence.

The bottom morphology

In chart No. 12 a 3-m depth contour runs along the shore at a distance of 150–250 m in the immediate vicinity of the harbour, and a 6-m contour 300 m from shore, except at the western flank, where the distance is up to 500 m. The depth figures further offshore are too scattered to give a more precise view of the bottom topography.

Echo-soundings in sections perpendicular to the shore have shown that along the shore east of the harbour runs a nearshore shelf with a slight inclination and with its outer edge at a depth of 2.5–3.5 m. The bottom undulates on account of the existence of subaquatic bars parallel to the shore. The grain-size distribution of the bars has been specially investigated in a section along an oil-discharging pier in the eastern part of the bay (cf. Fig. 74). The pier has an open construction, and its effect on the wave motion is insignificant. In this section the shelf is followed by a continuous gentle slope, but in the middle of the area east of the harbour (Fig. 62) there is also a plateau at a depth of 8 to 10 m. Immediately east of the harbour there is a shallow accumulation at the

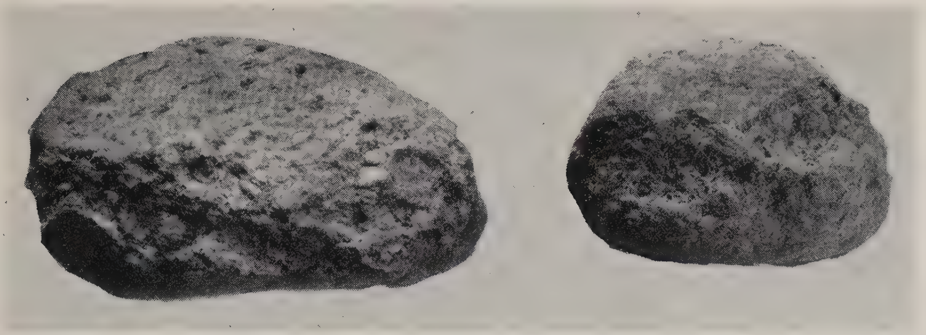


Fig. 58. Granite pebbles from a submerged beach ridge situated at a depth of 4 m west of the Jönköping harbour. The pebbles are viewed from the side, in the *in situ* position. The upper surfaces which were exposed to the water, and the lower ones which were in contact with gravel and coarse sand are hardly weathered at all. There is only a narrow weathered zone round the pebbles at the contact between the water and a thin mud cover.

breakwater (partly artificial?) and a narrow platform at 3.5 m.

The bottom morphology west of the harbour is of a different character. The nearshore shelf is narrow, and the slope is even down to a depth of 4–5 m, where there is a platform. In this a broad low ridge is found. Echo-soundings, made in 1961, indicated that this ridge contained soft material, probably of organic origin. Diving in 1962 verified this interpretation. Ridges with oblique angles to the shore, estimated to be orientated ENE.–WSW., and with a steeper slope towards the south, were examined. They were situated in front of a more coherent, low and broad ridge. The ridges partly consist of gravel. Between them and also under the coarse material greatly mouldered peat and stumps of wood were found.

The diverging ridges can be regarded as being cusplets. It seems probable that they were formed during the submergence, when a barrier was abraded and the beach moved in over a swamp. The orientation of the ridges indicates a transport towards the central part of the bay. That the pebbles have been at rest for a long time is obvious by the weathering which has affected them (Fig. 58).

These observations show that not all parts of the bottom of the bay have a sediment distribution that can be correlated to the present conditions.

The bottom morphology east and west of the bay

Along the western shore is a narrow nearshore shelf with a depth in the outer parts of about 2 m. From that the bottom slopes to a broader platform, over which the depth increases from 7 to 10 m. This platform abruptly ends with a steep slope (cf. Figs. 144 and 145).

To the east the bay is flanked by the severely abraded, glacial and glacifluvial deposits of the Rosenlunds bankar. The part nearest to the bay has the same profile as along the pier. There is a shallow nearshore platform, followed by an even offshore slope. Along the outer part of the abrasional platform the slope falls in steps.

Sediment sources

The main source for material transported into the bay is the Rosenlund cliffs east of the bay. The annual volume eroded in this area at the present time has been estimated at 16,000 m^3 (cf. p. 192). From underwater observations it is certain that almost none of this moves eastwards into the Huskvarna bay. Along the western shore north of the Jönköping bay the annually eroded volume has been estimated at 3,000 m^3. About one-third of the eroded material is coarser than fine sand, and less than 1 % coarser than sand.

If it is assumed that 16,000 m^3 enters the bay from the east and this volume is distributed over a width of 500 m and length of 2,500 m, the average annual accretion is 13 mm. Because of the composition of the material it will not be uniformly distributed, but the figure indicates that the amount is more than enough for the shore profile to keep pace with the submergence. With regard to the source and the sediment distribution on the shelf of the

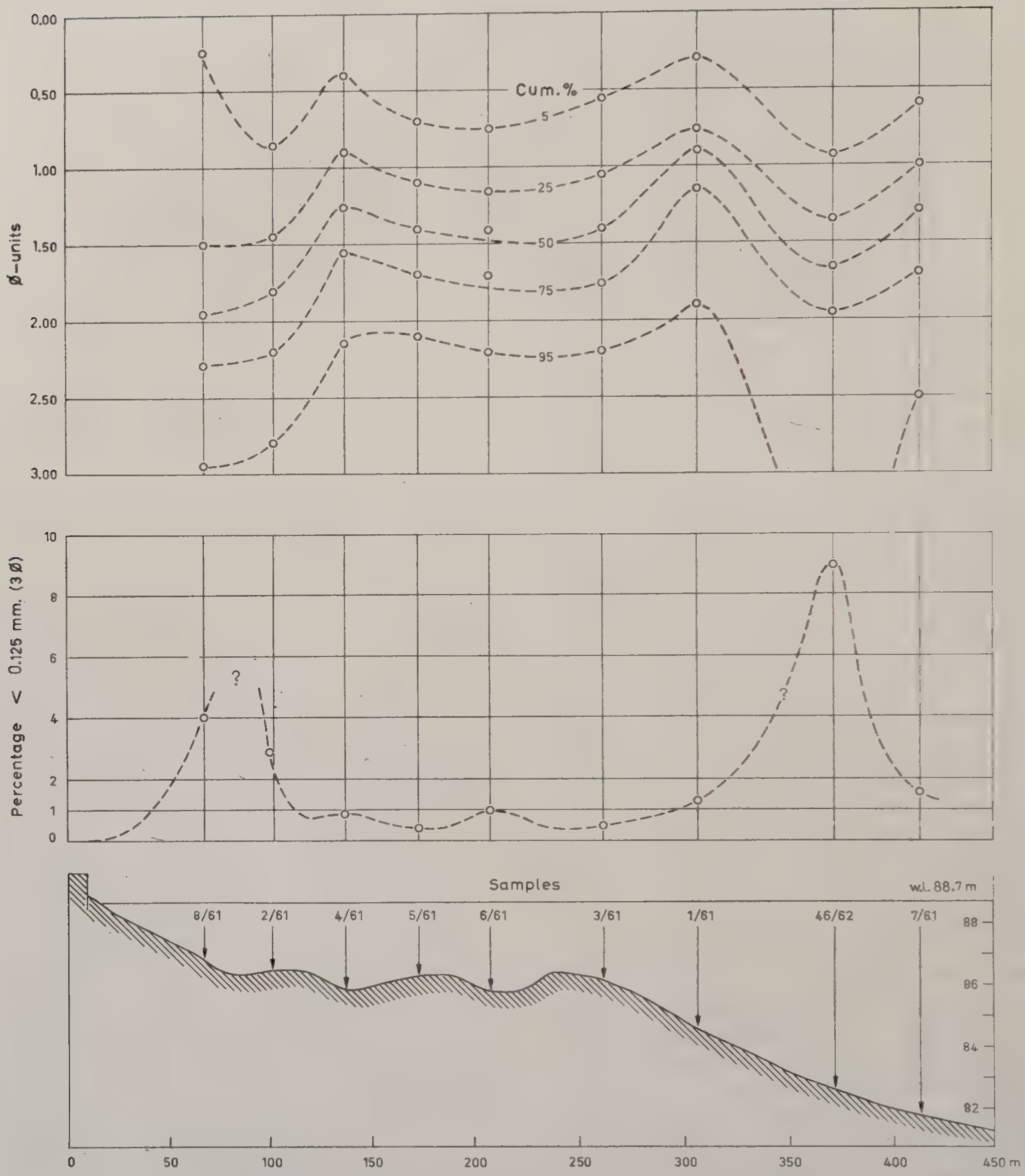


Fig. 59. Grain-size distribution in the nearshore zone of the pier section in the Jönköping bay.

bay, it has been estimated that the annual accretion in the nearshore zone is at most 4 mm, of which 1 to 1.5 mm is necessary to keep the bottom level in a constant relation to the rising water level. Off the shelf the accretion can be expected to be at least 10 mm.

The sediment distribution along and off the pier

The nearshore profile with its three subaquatic bars has already been discussed. In the summer of 1961 samples were taken on the crests and in the troughs. The grain-size distribution is

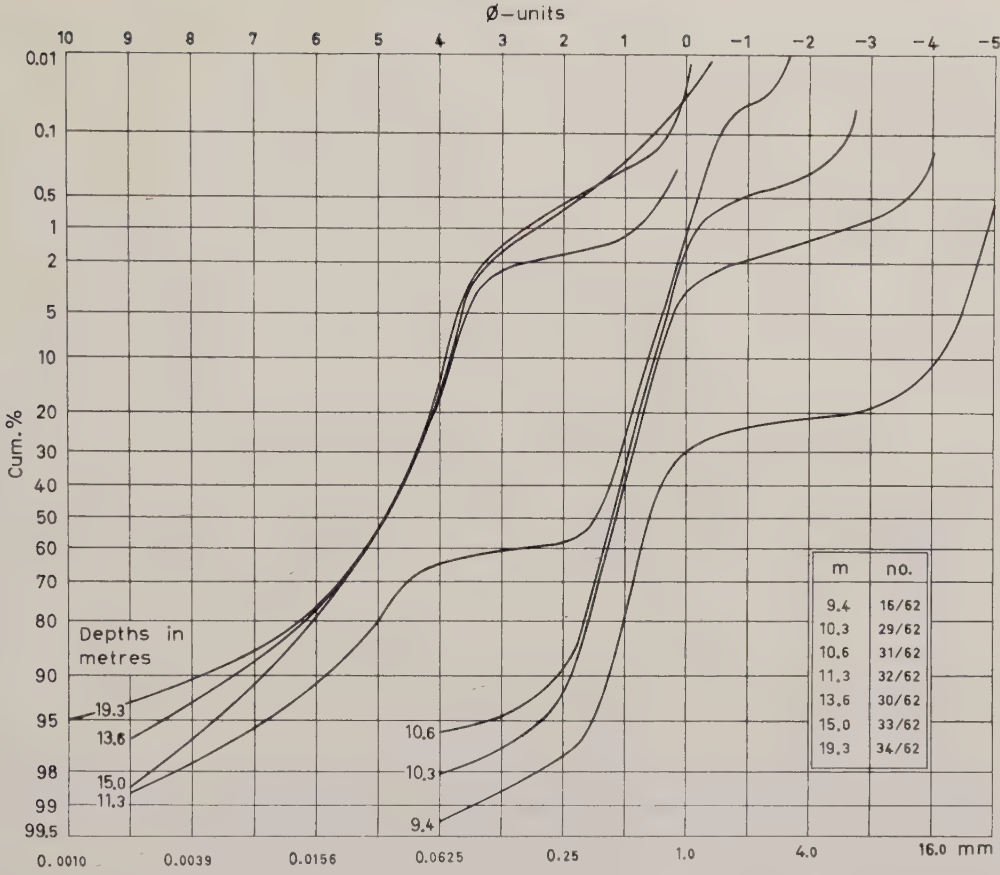


Fig. 60. Grain-size distributions of samples from depths of 9.4 to 19.3 m in the pier section, Jönköping bay.

illustrated in the upper diagram of Fig. 59. Medium sand is the characteristic fraction. The sorting is good. Except for the most shoreward sample, the distributions are unimodal and log-normal within the 2–98 % range. There seems to be a systematic variation of the grain size over the shelf, with the coarsest material on the front slope of the outer bar and on the front of the inner bar. The part finer than 0.125 mm is slightly higher in the outer troughs than on the crests, and considerably higher in the inner trough. This distribution probably reflects a sedimentation caused by small summer waves. The slope of the inner breaker zone consists of coarse sand with a top layer of granules and pebbles.

The zone off the shelf was investigated in the summer of 1962. The upper part of the slope down to a depth of 5 m has an average gradient of 1: 29.5; below that the even slope has a gradient of only 1: 75 down to a depth of 40 m.

From the investigation of the shelf in 1961 a decrease in grain size and a transition to the distributions characterized by suspended load was expected. However, even in the field it could be observed that the samples were sandy, even at depths below 10 m. In some samples very thin strata of coarser particles were visible, especially in sample 32/62 taken at 11.3 m. These strata were not really coherent, but were marked by single coarser particles at specific levels.

The grain-size distributions of samples taken in this section between 10.3 and 19.3 m are shown in Fig. 60. One sample, taken at 9.4 m west of the section, has been added. The diagram shows a consistent transformation of the distributions with depth. At 9.4 m there is an extremely well-marked bimodality. About

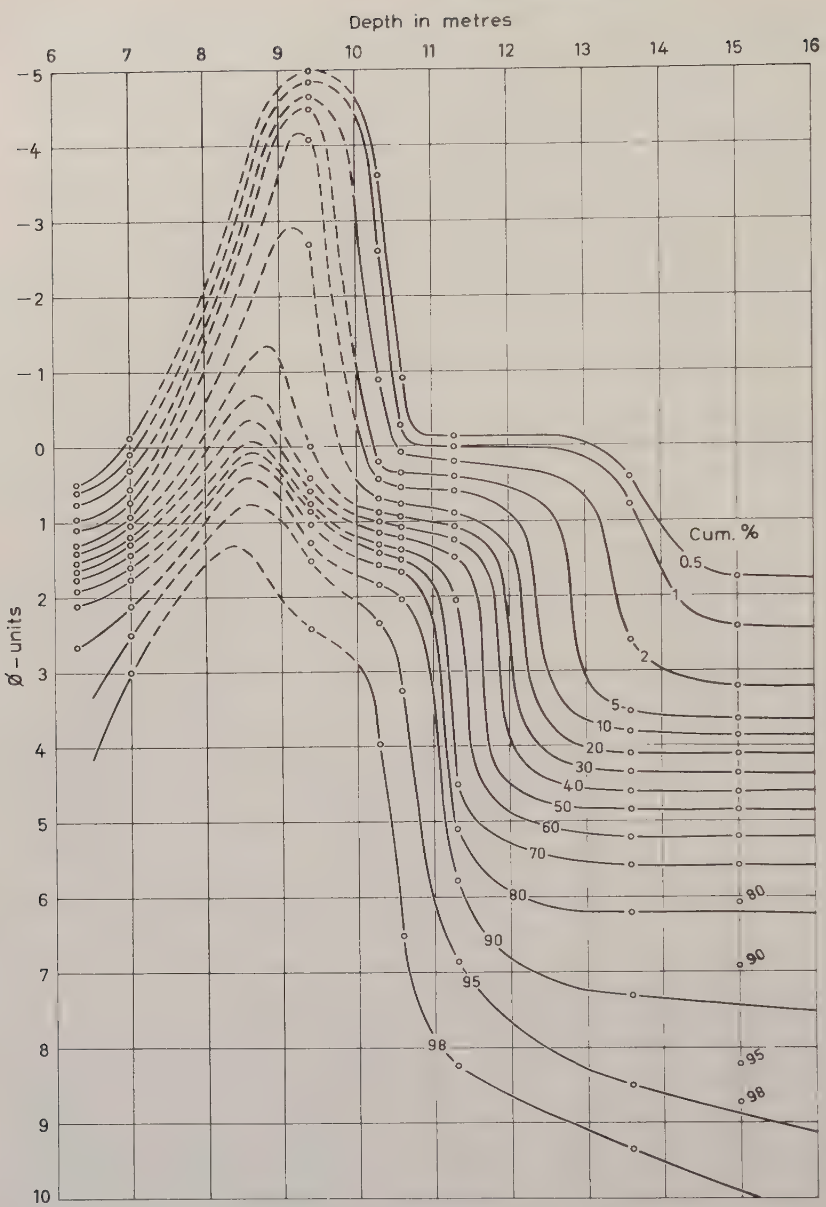


Fig. 61. The variation of grain size between depths of 6.3 to 16 m in the pier section, Jönköping bay.

20 % is coarser than –3 Ø, between –3 and 0 Ø there is only 10 %, the main part, 65 %, lies within the range 0–1.5 Ø, and finally there is a small tail of fine and very fine sand. Thus the main constituents are well-sorted coarse-to-medium sand and pebbles.

With increasing depth the particle size of the coarser fraction decreases and also its percentual part. Down to 10.6 m there is a small increase from 65 to 71 % in the interval 0–1.5 Ø, and the finer part of the medium sand and the fine sand (1.5–3.0 Ø) increase from 5 to 24 %. The sorting at 10.6 m is very good ($\sigma = 0.45$), and the distribution is strictly log-normal within the 0.5–84 % range (0–1.5 Ø). The amount of material finer than sand is still small (< 5 %).

Between 10.6 and 13.6 m the fundamental transition takes place. The amount of material finer than 3 Ø changes from less than 5 to

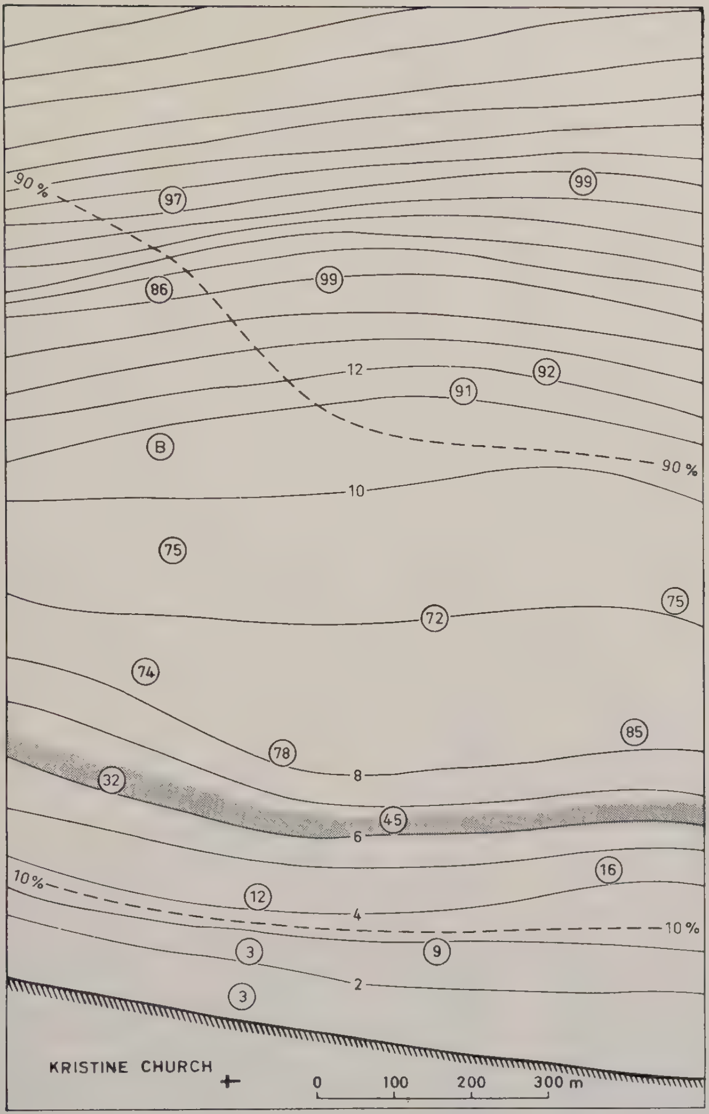


Fig. 62. The central part of the Jönköping bay. Circles mark the site of bottom samples, and the figures the percentages of material finer than 3 Ø. The approximate zone of transition from bed-load sediments to suspended-load sediments is shaded.

98 %. As can be judged from the sample taken at 11.3 m, there is no supply of fine sand and very fine sand (1.5–3.5 Ø) in the depth interval of transition, which results in bimodal distribution.

From 13.6 m to 19.3 m there are no significant changes. An analysis (35/62) of a sample from 40 cm below the bottom surface at 19.3 m shows a distribution almost identical with that of the top layer. These silty sediments have skewed distributions with α about 0.3 and a sorting coefficient σ of 1.2 to 1.3.

The grain-size distribution in relation to the depth of the interval from 6 to 16 m is illustrated in Fig. 61, which is thus connected with the diagram in Fig. 59. Because the sample from 9.4 m was not taken in the section, the curves of the interval from 7 to 10 m are dashed, but the distributions at 6.4 and 7.0 m and the distributions at 10.3 and 10.6 m indicate the existence of a peak corresponding to the 9.4 m sample. The transition zone in the interval between 10.6 m and 13.6 m is marked by the vertical part of the curves, which rolls up from 7 to 3.5 Ø.

The sediments of the lower part of the slope

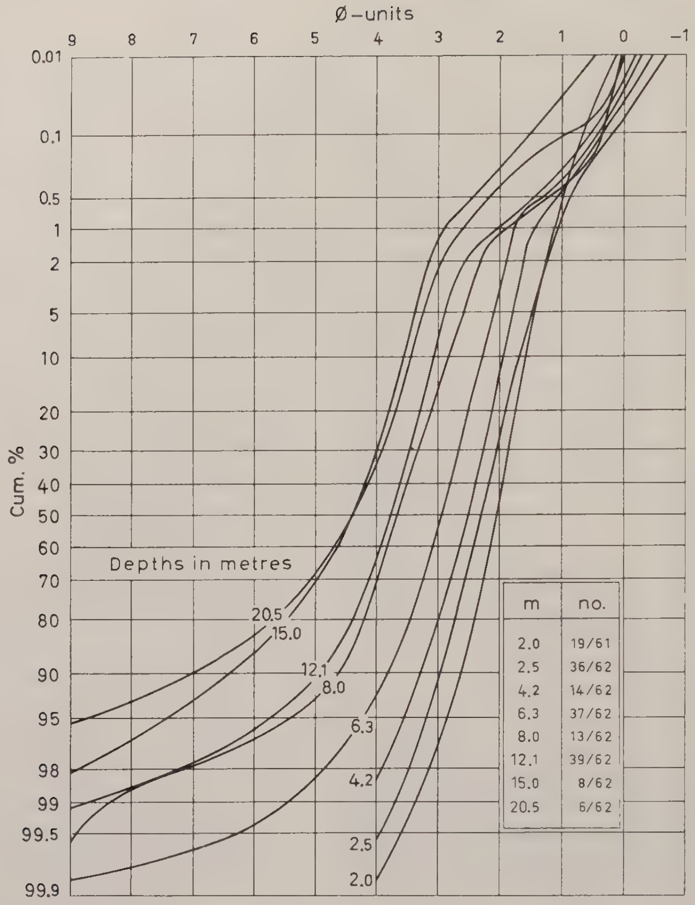


Fig. 63. Grain-size distributions of samples from depths of 2.0 to 20.5 m in the central part of the Jönköping bay.

have a type of distribution that corresponds to Inman's slope type, even though the Vättern sediments are somewhat finer. The sediments of the upper slope are about 3 phi units coarser than Inman's shelf sediment, and the sediments of the subaquatic bar zone are also much coarser than any of his sediments.

The lack of material in the range 1.5–3.5 Ø, which causes the bimodality, could be explained by lack of supply. However, there is much fine sand in the main source–the Rosenlund area–and when this sand, the easiest-moved fraction, is not found in the nearby profile, this can only be explained as an effect of longshore drift, which in the breaker zone has moved this fraction into the central part of the bay.

The process which has deposited the coarse fraction in the upper slope cannot be reconstructed from the sediment distribution. From comparison of the transport diagrams Fig. 42, the null-point diagrams Fig. 41, the wave-height diagram Fig. 35 and the energy diagram (Fig. 56), transport by direct wave action seems possible in periods of the strongest winds. The main fraction of the samples in the 9.4–11.3 m interval roughly corresponds to the expected particle size under steep waves in the upper part of the energy maximum.

The central part, east of the harbour

In this area 20 samples have been hauled from depths between 1.7 and 20.5 m (Fig. 62). There is a nearshore shelf with low bars and its outer edge is situated at a depth of 2.5 m. In the offshore slope a broad flat step is found between 8 and 11 m. Of these samples only one (4/62) from 10.3 m has a bimodal distribution. The point of inflexion lies at 1.0 Ø.

In order to illustrate the relation between grain size and depth, the distributions of eight

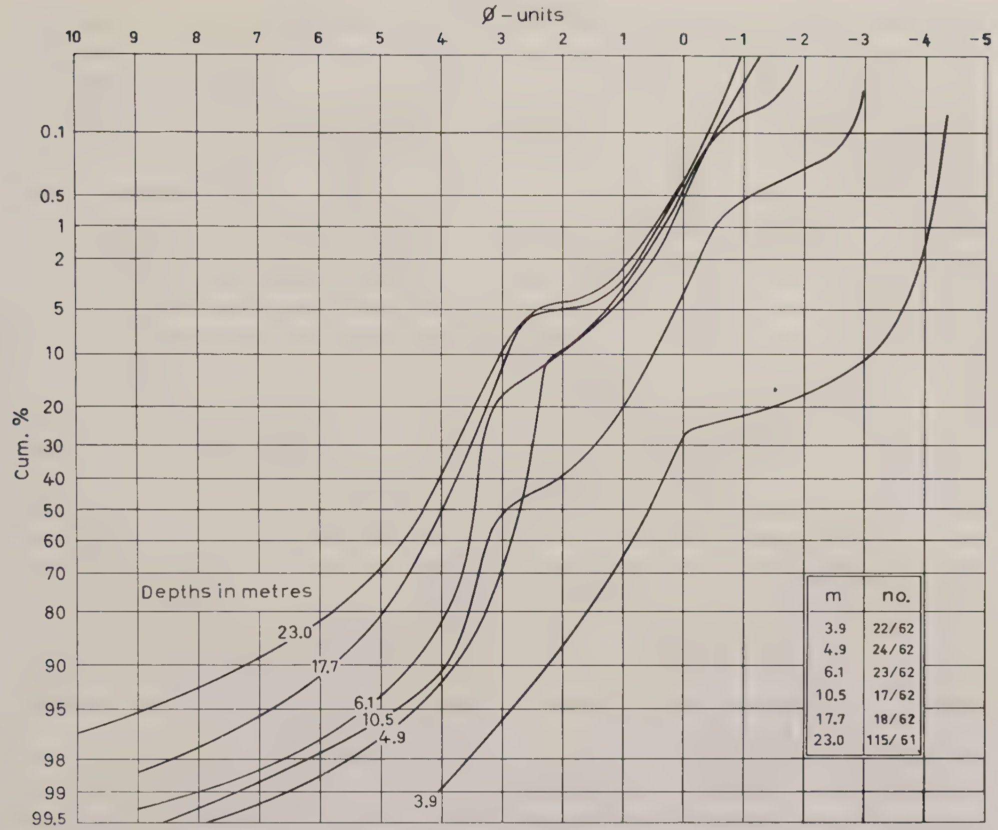


Fig. 64. Grain-size distributions of samples from the western section of the Jönköping bay.

samples from depths between 2 m and 20.5 m are shown in Fig. 63. The other samples also fit the system, but more curves would have made the diagram too crowded. As can be seen, less than 1 % is coarser than 1Ø at all depths. With increasing depth the sediment constantly becomes finer. Through the increase of fine material the distributions become more skewed, and the rising tail, seen in the diagram, also causes a decrease in sorting.

The distribution of the percentual part finer than 3 Ø is illustrated in Fig. 62. One finds that a rapid transition from less than 30 % to more than 70 % takes place in the lower part of the nearshore shelf slope, which thus marks a transition from bed load to suspended load.

When the distributions of this central area are compared with the distributions of the eastern section in Fig. 60, it is striking to note that they fit exactly into the interval between 1 and 3 Ø, where there are extremely small amounts in the eastern section. Thus, the sediment pattern can be explained by longshore drift of this material towards the central part of the bay.

The samples below the shelf slope are of the same type as Inman's slope sediments; the upper samples are up to 1 phi unit coarser than his typical shelf distribution, but the sorting is identical.

The area west of the harbour

From the finds of submerged peat and beach ridges the risk of contamination of deposits referable to the present and earlier conditions is obvious. This risk can be estimated to be smallest on the shallow shelf, where the bottom is certainly directly affected by the present wave action, and in the actual offshore parts, where deposition of suspended material can be expected to have predominated for a long time. As we found in the previous section, the variation in grain size is small at great depths.

Some representative curves have been drawn in Fig. 64. Sample 22/62 was taken inshore of the area with peat 600 m WNW. of the harbour, the other samples straight offshore, except sample 24/62, which was taken 800 m north of the section.

The sample 22/62 from 3.9 m has a distribution which greatly resembles that from 9.4 m, taken west of the pier section. The coarse material is of the same grain size, and the bimodal gap has the same range with a break at 0 Ø. In their bimodality the other curves are also like those from the area off the pier. With respect tó the grain size they fit into the transition zone between 10.6 and 13.6 m, but they are not well correlated with regard to depth.

Sample 115/61 from 23 m is the surface material of a core with a length of 45 cm. The sediment has a homogeneous greyish appearance, except for some very thin blackish strata, coloured by organic substances. The position of these strata is shown in Fig. 65. The interval between them varies from 4 to 54 mm. Between 11.5 and 15 cm from the top there is a broader but not so dark band.

At the four levels analysed (115–118/61) the sediment has an identical type of distribution, and the absolute variations are small (Fig. 65). The general conditions can be regarded as having been the same during the deposition of these upper 45 cm.

The origin of the organic layers is suspicious. The rather regular frequency gives the impression of a seasonal dependency. The sharp limits indicate that the deposition of the organic substance takes place under very specific conditions, which are either fulfilled or not. It should be remarked that these layers were not found in the core taken at a depth of 19.3 m off the pier.

The most extreme seasonal change in the conditions of sedimentation is connected with the formation of lake ice. When the ground is frozen and the shores are covered with ice, the supply of new material is stopped, but transport is still possible from the nearshore area. When the lake becomes capped with ice, even this action ends. Under the ice, fine organic particles can still move in suspension, and in the vicinity of the harbour there is a flow of organic material from Lake Munksjön and from drainpipes.

If the formation of the layers is exclusively bound to the conditions of the frozen lake, a run of years without ice cover could be expected to be marked by a larger interval in the sequence. However, the variation in yearly sedimentation is unknown, and we do not know if the organic material is sometimes eroded. An attempt was made to correlate the series of winters with an ice cover and the sequence of the core, but no certain correlation was found. However, it was found to be most probable that the large interval 19 cm below the surface corresponds to the ice-free period 1942–1945, which gives an average rate of sedimentation of about 1 cm/year. This figure seems reasonable with regard to the estimated amount of material transported to the bay.

Why has the sediment of this core and also the other samples from west of the harbour bimodal distributions with a deficit of fine sand, and why have not the submerged beach ridges etc., found at depths of 4 to 5 m, been covered with sand? The eastern shore of the bay consists of a barrier or bay-mouth bar built out to the outlet from Lake Munksjön, where the harbour is situated. Immediately west of the harbour the breadth of the nearshore shelf is only 100 m, but in the most westerly part of the bay it is 500 m. It can be estimated that the predominant drift in the bay has been to the west. The sand, brought along the shore from the east, settled on the shelf of the barrier and at the end of it, and only the suspended fine material could be carried further westwards. By the construction of the harbour, the drift on the shelf has been totally stopped. Thus it seems probable that there has been a constant lack of supply of sand near to the west of the outlet from Lake Munksjön. East of the harbour there is no advance of the shore, because the material on the shelf is too fine for the steep waves.

Conclusion

By direct wave action a shelf has been built out in a depth of 2 to 3 m. The sediment is a well-sorted, log-normal-distributed medium sand. The sand is somewhat finer in the central part of the bay than in the updrift area.

On the slope off the shelf of the central part of the bay there is a transition, in depths between 6 and 8 m, from sediments with less than 30 % finer than 3 Ø to sediments with

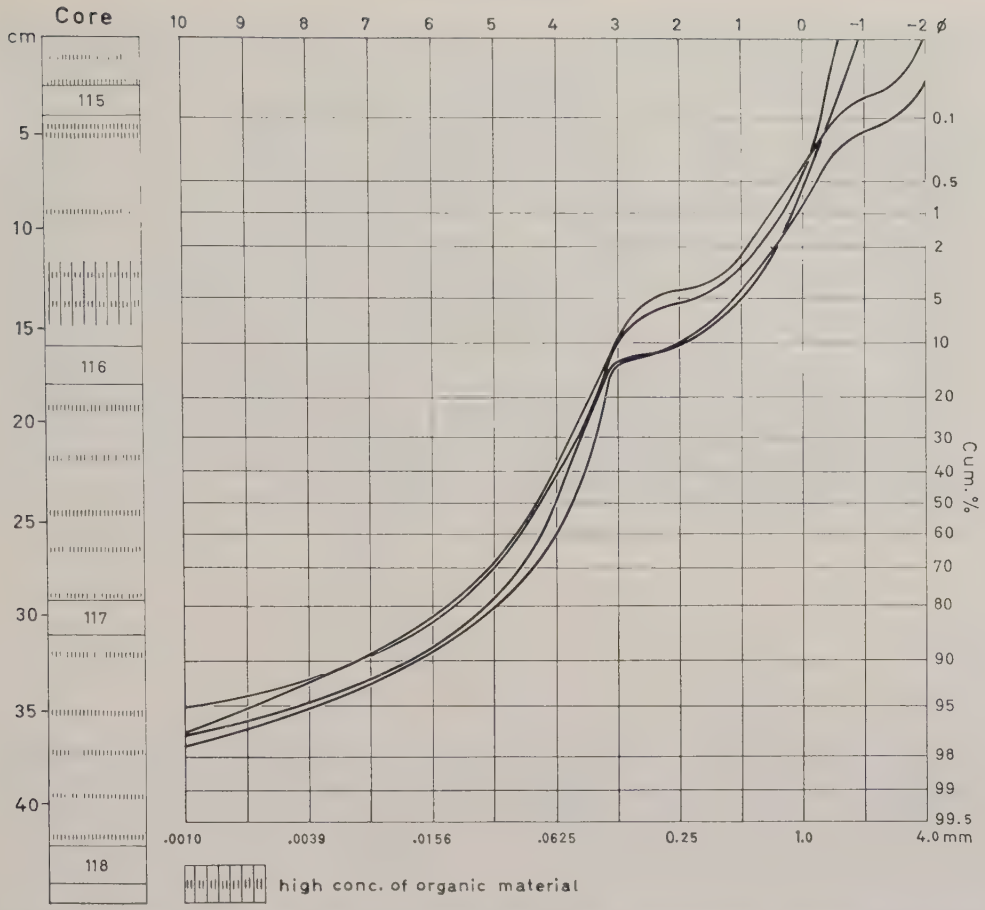


Fig. 65. Bottom core from the outer end of the western section of the Jönköping bay. Water depth 23.0 m. Grain-size distributions at four levels in the core (nos. 115—118/61).

more than 70 % finer than 3 Ø. The former sediments have log-normal and well-sorted distributions, while the distributions of the latter are skewed by an increasing amount of silt. On the slope of the updrift area fine sand and very fine sand are not found, because this material is transported along the shore to the central part of the bay. This sorting activity causes the distributions to be bimodal in the transition zone of the eastern updrift area. The same bimodality is found in the western part of the bay. The transition to sediments predominantly consisting of particles finer than 3 Ø is in these areas not concluded until depths of 10 to 12 m are reached.

Wave action in plan and shoreline configuration

The directional resultant of wind work of Visingsö

In the investigations on the Visingsö shores in the 1950s the DRW, according to Schou, was calculated on the wind statistics for the six years 1945–1950. Winds of a force of 4 Beaufort were included, because it was found that they generated waves with a transport capability. As could be expected from the wind-frequency diagram (Fig. 18), the polygon of the vector summation is elongated (Fig. 66). The S.–SE.–ESE. winds outbalance the

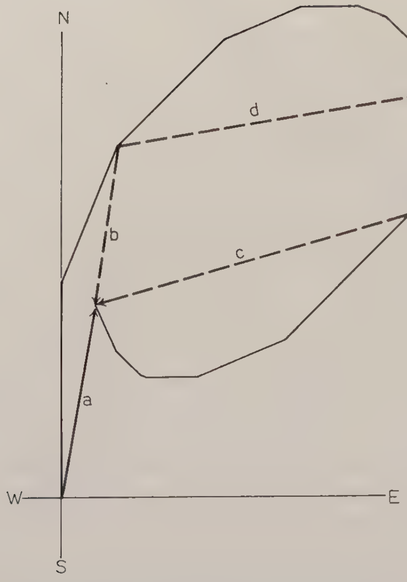


Fig. 66. The directional resultant of wind work (*a*) of Visingsö, calculated on the statistics of the period 1945—50, (*b*) the resultant, when winds from the S. and SSE. are omitted, (*c*) the resultant of winds affecting the eastern shore, and (*d*) the resultant of winds affecting the western shore.

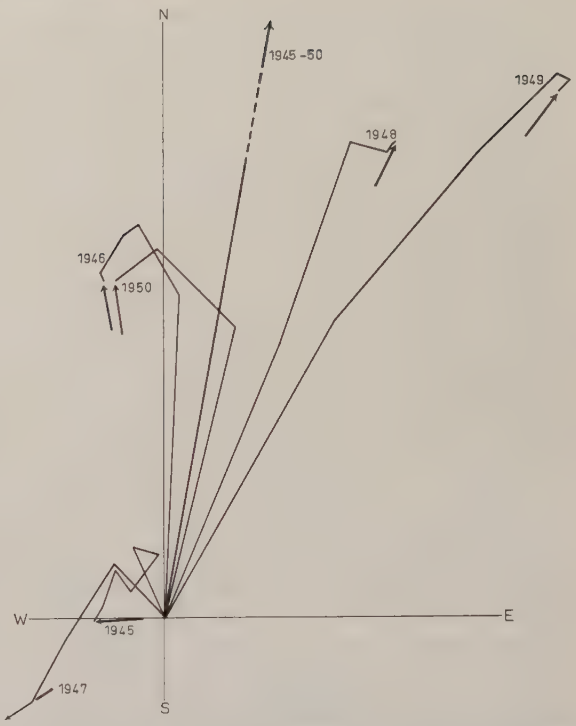


Fig. 67. The direction resultant of wind work of Visingsö for single years (arrows) based on vectorial summation of the resultants for each Beaufort unit, starting with 4 *B* at origo.

N.–NE.–ENE. winds, and the resultant (*a*) is directed northwards, parallel to the axis of the lake. It can be noted that, where the S. and SSE, winds are shadowed, the resultant will be almost exactly the reverse (*b*), and the same relation is found between the resultants of the eastern (*c*) and western shores (*d*).

In Fig. 67 the resultant for every single year is shown, constructed on vectorial summation of the resultants for each Beaufort unit, starting with the lowest (4 B). The construction clearly demonstrates two facts. Firstly, that the resultant varies considerably from year to year, but generally lies in the interval NNW.–NE., and secondly, that its direction mainly depends on the low velocities. Hence it is correlated to the predominant winds rather than the dominant winds. If the weak winds are excluded, the resultant will vary between easterly and westerly directions. This can be directly found from the wind statistics. The stronger winds are fairly balanced with regard to the axis of the lake, and small variations in their directions will cause a shift of the resultant from east to west. In waters with a fetch distribution like that of Lake Vättern the DRW can only be regarded as an explanatory illustration of the azimuthal wind-force conditions. No quantitative conclusions with respect to shore evolution can be directly based on the resultant.

The possibilities of beach-drifting and the direction of net movements

If there were no irregularities in the shoreline configuration and free beach-drifting was possible all along the shores, a directed drift could be expected to exist over long stretches towards the ends of the lake. In reality the beach-drifting is interrupted in many places. In the northern section of the lake the shoreline is too dissected for continuous drift, and rocks

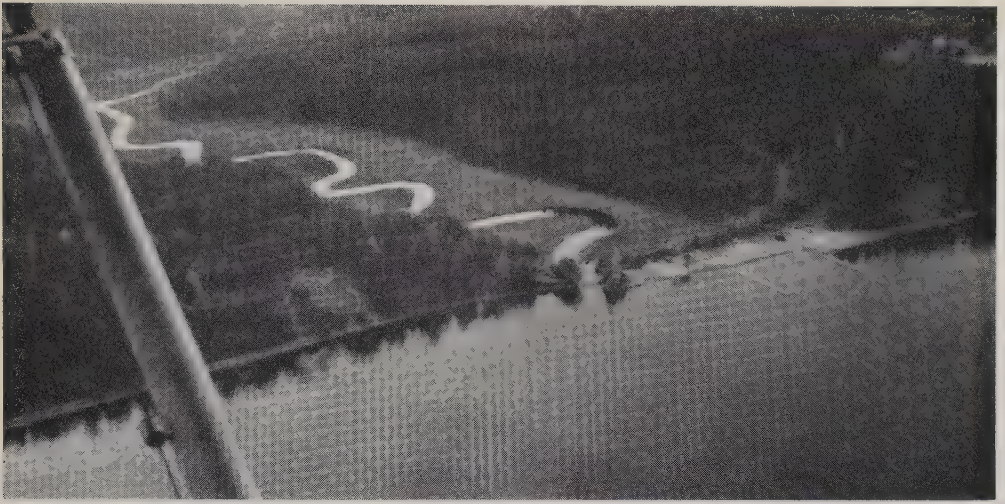


Fig. 68. The submerged valley of the river Dumneån in the Bankeryd area. The river meanders in peat deposits. The cliff with the erosion scar in the foreground has a height of 16 m. Note the barrier built out from the north, and the deposition north of the recently constructed jetties. Oct. 1962.

also frequently descend directly into deep water. In the Motala bay and the Karlsborg area the configuration also hinders a continuous drift. In the central and southern section beach-drifting cannot take place past Mount Omberg and the cliffs south of it down to the Gränna area. South of this area the drift is again stopped by steep rock cliffs. Along the western mainland shore beach-drifting is obstructed by rock outcrops at the latitude of the northern end of Visingsö island, and the cliff shore south-east of Bankeryd cannot be passed. Thus the stretches of possible free drift are very restricted. There are the shores of the narrow southernmost part of the lake, the Gränna area, the shore from Bankeryd to north of the Svedudden promontory, and the shore along the lowland of Guldkroken.

The net direction of beach-drifting can be traced by spits and barriers built out in the downdrift direction, by depositions on the updrift side of obstacles on the shore and erosion on their downdrift sides, by skewed distributions of beach material in bays, and by skewness in the form of indentations, and finally by decrease in particle size in the downdrift direction.

In the area north of Gränna harbour a southerly net-drift direction is indicated by deposition of shingle north of landing stages and outlets and lack of material south of them. South of the harbour there are no indications of a net drift, until the shoreline turns to a south-easterly direction, when the net movement is again southwards.

Along the undulating shoreline of Guldkroken no specific net-drift direction can be proved with certainty. In a shallow indentation west of the Flisen shoal there is more beach sand deposited in the northern part, but the dune ridge which fringes the shore is higher in the southern part. This distribution could be explained by stronger winds from the north-east but a greater fetch to the south. South of Hjo there is a tendency towards larger sand deposits in the southern parts of the indentations, and in the pocket beaches south of the Guldkroken area more shingle is accumulated in their southern parts.

On the Svedudden promontory a spit along the northern shore proves a southward drift (cf. Fig. 10). The Svedån river formerly flowed along the spit and had its outlet on the southern shore. Whether the present outlet is the result of a natural break or an artificial cut is uncertain. North of the present outlet a rock-fill jetty was built in 1960–61. There is some accretion north of this, but sand is quarried on the shore and material is dumped south of the outlet, so the drift cannot be quantitatively estimated by the effect of the jetty.

South of the promontory there are no certain traces of a net drift on the shores. In this area one of the very few localities, in the open lake, of sedge growing off the shoreline is found. Where the shore turns to a more north-southerly direction, there are more erosion scars in the north-facing slopes of points than in the south-facing. At the outlets of rivulets, which form deep ravines in the glacifluvial deposits, there is, as a rule, a spit of gravel and sand projecting from the northern shore.

In the Bankeryd area predominant beach-drifting towards the south-east is indicated by a spit at the outlet of the Dumneån river and by the deposition on bathing stages recently built at this place (Fig. 68). The shore has retreated very little during the last 200 years. That the balance of the drift is delicate is shown by the fact that the construction of a small landing stage 350 m south of the river immediately caused severe erosion on the downdrift side (cf. Fig. 106).

On the southernmost part of the western shore the drift is, of course, also to the south. There are many landing stages, and north of each there is accumulation and south of each a tendency to erosion.

The sediment distribution in the Jönköping bay has already been discussed. It will only be repeated that the part east of the harbour is a barrier beach built out by beach-drifting from Rosenlunds bankar. In the Huskvarna bay a spit has been built out from the west at the mouth of the Huskvarnaån river. Along the eastern shore, north of the Huskvarna bay, there is no beach-drifting of importance, because the shore mainly consists of residual boulders and blocks.

It can be concluded that in the middle part of the lake the direction of beach-drifting is rather unspecific, but from the latitude of the Gränna area it is generally directed towards the south. However, south of the Gränna harbour and the Svedudden promontory there are no traces of a specific net drift. This problem will be discussed in the next section. In the northern part of the lake the shores are too dissected for a continuous drift.

The conditions in areas of wave-energy deficit

Areas of wave-energy deficit

The potential wave heights generated by winds blowing along the lake are much the same at the Jungfrun skerry, the Flisen shoal, the northern end of Visingsö and in the Jönköping bay (cf. Fig. 35). The small variation is a function of the limiting effect of the width of the lake on the effective fetch. The wave heights of the southern end of Visingsö are lower because of shorter shore-to-shore fetch lengths and the narrowness of the lake, south of Visingsö. Low wave energy can be expected along the middle part of the eastern shore of Visingsö and the south-western shore of the island, also in the vicinity of Gränna harbour, in the area south of the Svedudden promontory and along the eastern mainland shore, north of Huskvarna. The reasons, as regards the Visingsö shores, are their orientation and their nearness to the mainland. As regards the Gränna and Svedudden areas, there is a sheltering effect of Visingsö island. The shore, north of Huskvarna, is situated in the narrowest part of the lake and deviates more than 90° from the direction of the lake.

In areas of low energy surrounded by areas of higher energy, accretion of sediments can be expected (cf. Tanner 1961). In this respect the area north of Huskvarna is of small interest. As it is situated at the end of the lake, the transport will certainly be towards the Huskvarna bay.

The Gränna area

The moraine cliff of the Gränna area is being actively eroded. With respect to the intensity of the retreat it cannot be compared with the shores exposed to higher wave energies to the north and the south, because they consist of bedrock cliffs. The beach has a width of 2 to 4 m, and is made up of sparse shingle and residual boulders. The bottom conditions are of great interest, because there is a sanding up in depths between 1 and 5 m around the harbour, and also off the moles at the entrance to the harbour. In order to keep the entrance clear, two jetties have been built south of the harbour. Their effect is uncertain. There is no beach accumulation on the windward side, south of the jetties, and thus there is at least no northwards-directed beach-drifting. On the contrary, shingle is accumulated on the shore end of the northern mole of the harbour (where there is no jetty). The fact that there is no sand on the shores, though there is on the nearshore bottom, can be regarded as the common effect of steep waves. The shingle accumulated north

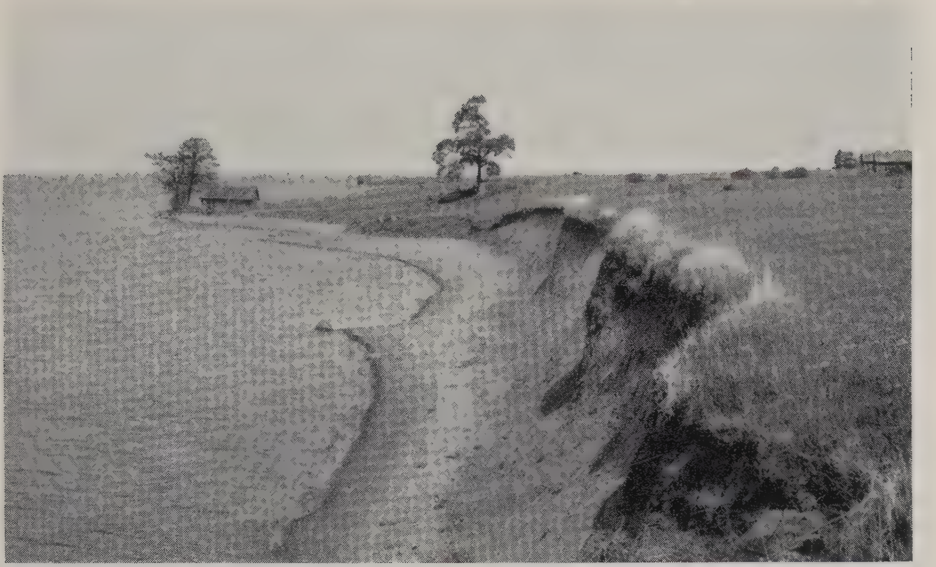


Fig. 69. The small ness at Rönäs on the western shore of Visingsö island, viewed from the south. The old alders on the shoreline fell in the following year. Aug. 1962.

of the harbour proves that the stronger winds and the longer fetches from the north are of greater importance to the transport of coarse material than the more frequent weaker winds from the south. It is not known if the construction of the jetties south of the harbour was based on measurements of sand transport in the nearshore zone or on estimations of the effect on sand transport of the more frequent waves from the south-east. If there really is a northwards-directed sand transport, the net movements of shingle on the shore and sand in the nearshore zone are in reverse directions. The sanding up in this low energy area is at least certain.

Visingsö island

Along the middle part of the eastern shore of Visingsö the shore-process activity is low. The beach has only an average width of 1 to 2 m, and consists of shingle and residual boulders. North of Visingsö harbour there is almost no erosion. The moraine cliff is covered with grass, alder trees and bushes. South of the harbour there are some minor erosion scars and outcrops of shale near the shoreline. The cliff is forested with spruce and fir. Some sand is found on the nearshore bottom off this eastern shore, but there are no significant accretions.

The conditions of the western shore of the island are more complicated. The middle of the western shore is rather straight and runs N. 32° E., but 3.7 km from the southern end of the island (G in Pl. 1) the direction turns to N. 19° E. On the stretch between 400 and 700 m south of the turning point there is a beach accumulation of coarse sand with a maximum width (1962) of 28 m. Except for the Sandudden ness, this is the only accumulation projecting from the cliff line on Visingsö, and the only one in the whole lake projecting from an open straight cliff shore (Fig. 69).

North of the accumulation area the shore consists of a moraine cliff and a narrow beach with a width of 1 to 3 m. The material of the beach mainly consists of residual and semiresidual material. The average shore retreat has been between 8 and 12 m during the last 100 years. South of the accumulation area the shore consists of a moraine cliff that has been stable for a long time, because it was protected by a shingle beach (cf. Fig. 70). The shingle reaches its average size of 3 to 4 cm 1,300 m south of the sand accumulation. South of this point the shingle dimension is fairly constant, but in

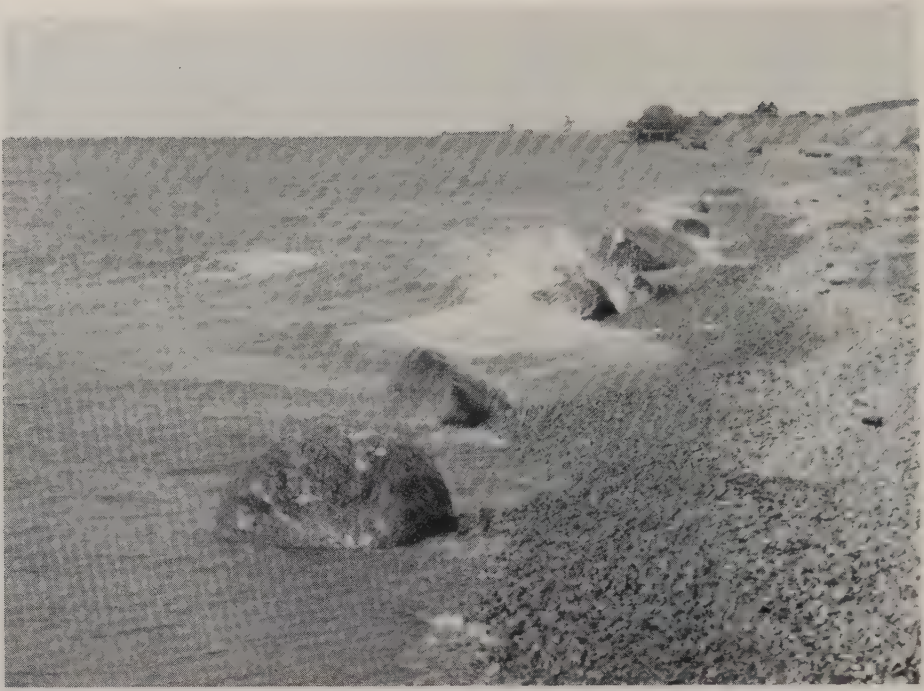


Fig. 70. From the shingle beach on the south-western shore of Visingsö island. A row of ice-pushed boulders. The trees of Fig. 69 may be seen in the far background. Aug. 1962.

some places near the southern end the shingle may be considerably coarser (mean size up to 6 cm). At artificial obstacles (launching rails for boats and boat-houses) there is regularly more shingle accumulated on the northern side.

An old topographic description (Allvin 1859, p. 23) mentions that in the area of the present sand shore there was an accumulation of material finer than "the usual shingle". The beach is said to have a width of 9 to 12 m and a length of more than 300 m, and it was noted that the sand shifted in position with the prevailing winds.

By carbon-14 dating it has been possible to prove that the cliff immediately south of the present accumulation has been stable since the 15th century (cf. p. 188). The sand accumulation began to be severely abraded in the early 1950s, and when its southern part had gone, a rapid abrasion of the cliff that had been stable for so long started in 1957. Why the loss of sand started just there is uncertain. It is only known that sand and gravel have been taken for roads in this area, and this may possibly have disturbed the drift balance. In this connection it may be mentioned that the shingle must have been transported from the north to a shore with less inclination to the northern winds blowing along the lake, and there the transport is balanced by the more frequent but lower waves generated over the shorter fetches from SSE.–E. The finer material has been deposited at the northern end of this shore, because it can be moved by the weaker winds from the south, but cannot be transported north of the turning-point of the shoreline, because the wave energy associated with the northerly winds increases north of this point.

The area south of Svedudden

In the area south of the Svedudden promontory there are no significant beach accumulations but an unusually broad shallow platform (cf. Figs. 71 and 72). The shoreline runs N. 24° E., while immediately to the south the direction is N. 10° W, and north of the Bankeryd area (north of E. in Pl. 1) it is N. 12° E. Thus the shore is less exposed to the northerly winds blowing along the lake than farther south. On the platform there are shoals of sand far off the shore. The characteristics of the sediment are given in Table 14.

As was mentioned in the introductory chapter on the geology of the basin, the foundation of this platform is probably sedimentary rocks and abraded glacial deposits, but the shoals of sand could not exist, if there was no accumulation in this area. A directed net drift would sweep away the sand lying within breaker depths. The sand of the shallowest part of the sections (nos. 63 and 64/62) has the same grain-size distribution as the sand of the subaquatic bar zone in the Jönköping bay. On the deeper southerly section the sediments are finer. It may be noted that in the area at the foot of the inner shore slope there are finer sediments, which can be regarded as having been deposited by direct outwash from the local beach.

The sediments of this area are also comparable with the sediments of the shelf off the Sandudden ness on Visingsö (cf. Fig. 2). In this exposed position the whole shelf is characterized by medium sand (Table 00, nos. 10–11/57) that has also been transported out over the edge of the shelf (no. 31/57) to the slope built up to the angle of static repose.

Table 14. Characteristics of sediments from sections south of the Svedudden promontory (62—66/62 and 67—69/62), the platform east of the Sandudden ness (10—31/57), and a section in the Bankeryd area (58—61/62). Phi units.

Nos.	Depth (m)	Md	σ	α	% <3 Ø
62/62	3.3	2.45	0.50	—0.10	11
63/62	2.3	1.55	0.53	—0.06	1
64/62	2.1	1.50	0.35	0.00	0
65/62	9.7	2.10	1.10	0.09	23
66/62	12.2	2.60	1.38	—0.12	39
67/62	1.2	2.10	0.45	0.00	4
68/62	4.0	3.45	0.58	—0.47	80
69/62	5.2	2.50	0.65	0.08	23
10/57	2.7	1.50	0.70	—0.43	0
11/57	4.0	1.65	0.38	—0.07	0
31/57	12.0	1.75	0.50	0.00	2
58/62	4.5	2.10	0.60	0.00	7
59/62	6.0	0.80	0.85	0.29	7
60/62	8.1	3.20	0.28	—0.07	79
61/62	10.7	2.95	0.50	—0.10	44

A comparison with samples from a section situated immediately north of the outlet of the river Dumneån in the Bankeryd area may finally be included (Table 14, nos. 58–61/62). There is no broad shallow platform in this area (cf. Fig. 73). In the mid-slope of the narrow platform ($d = 4.5$ m) the material consists of strictly log-normal distributed medium sand $Md_{Ø} = 2.10 = 0.23$ mm), but at the foot of this slope ($d = 6.0$ m) there is coarse sand $Md_{Ø} = 0.80 = 0.57$ mm). The distribution is truncated at –0.5 Ø (14 mm). This sediment formed a thin layer directly on a clay surface. In the deeper low slope the sediment is of suspended-load character. It is well sorted, but in the upper tail of the distributions there is a deficit with a point of inflexion at 2 Ø, that is, in the same interval of deficit found in the samples from the Jönköping bay.

The thin layer of coarse sand found at the foot of the platform slope proves that there is no net accumulation of bed-load sediments in this area, and that the depth is too shallow for a deposition of suspended-load sediments, but in a depth of 8 m this is possible. This depth of transition to suspended-load sediments corresponds to the depth found in the central part of the Jönköping bay. The coarse sand at the foot of the slope may have been deposited during previous time with a lower water level.

Shoreline configuration and orientation

Introduction

The shorelines, which have a configuration and orientation that can be related to wave-energy distribution (wind-frequency and fetch distribution) are developed by the abrasion of moraine and glacifluvial deposits and to some extent of sedimentary rocks. The narrow beaches are of small importance, as regards the horizontal configuration. Only in a few bays and at the outlets of some small rivers can this configuration be regarded as having been formed by a depositional process. In the main part of the lake the position of the present shoreline is the result of abrasional processes and submergence, which have caused a considerable retreat (cf. Fig. 126). Because depositional features are generally of no importance, the effect of inhomogeneity with regard to erosional resistance has not been blotted out. Some large features of shoreline orientation, illustrated in Fig. 74, will first be discussed, and then different types of configurations. The exposure is represented in the sketch maps by the interval of long shore-to-

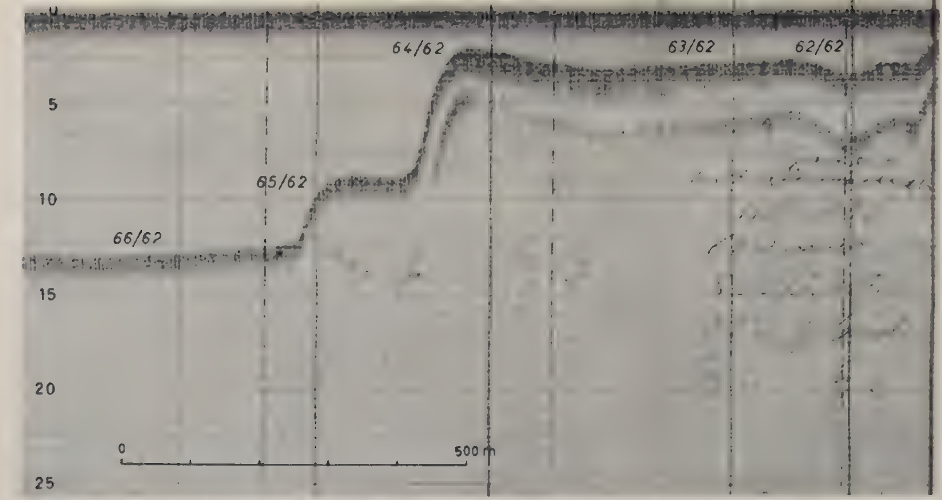


Fig. 71. Echogram of the bottom profile 3 500 m south of the Svedudden promontory. Direction N. 110° E. The profile ends 20 m from the shoreline. The site of samples 62—66/62marked by vertical lines.

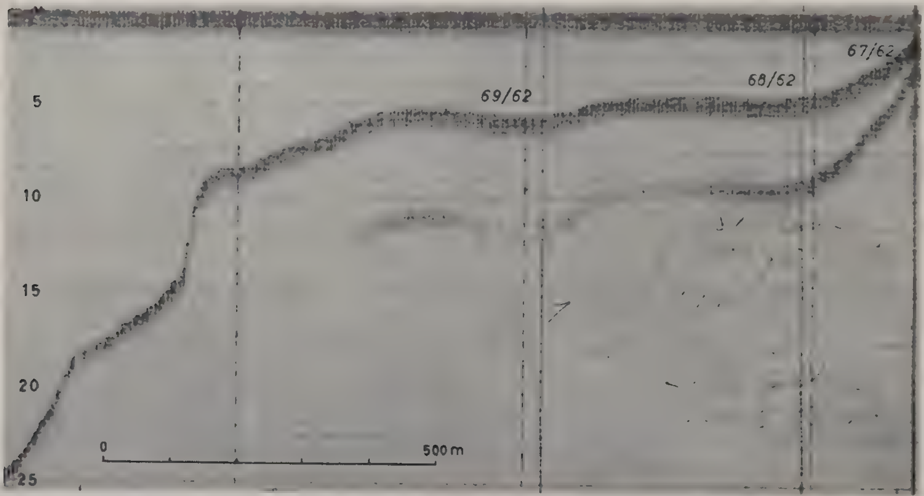


Fig. 72. Echogram of the bottom profile 5 000 m south of the Svedudden promontory. Direction N. 118° E. The profile ends 25 m from the shoreline. The site of samples 67—69/62 marked by vertical lines.

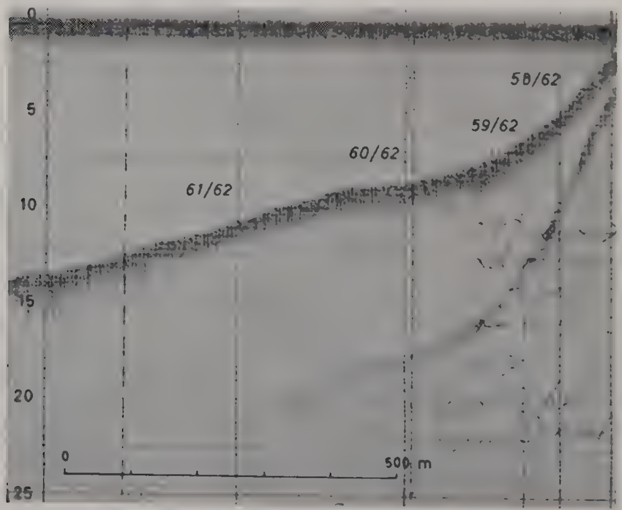


Fig. 73. Echogram of the bottom profile from the southern bathing stage at the mouth of the river Dumneån in the Bankeryd area. Direction N. 50° E. The site of samples 58—61/62 marked by vertical lines.

shore fetches and the direction of the maximum fetch. The shoreline orientation is illustrated by the perpendicular to the shore.

Bays

The eastern and central part of the Jönköping bay is a bay-mouth barrier built up of material abraded along Rosenlunds bankar. The western part is cut in shore sediments of the Baltic Ice Lake stage. The orientation of the shore has not changed in recent time. The perpendicular of the central part is 9° west of the maximum fetch, and it coincides with the axis of symmetry of the potential wave-height distribution (Fig. 35). The full width of the bay corresponds to a radius of circular curvature of about 5 km, but the central part is too flat to fit this circle.

The main part of the shore of the Huskvarna bay is abrasional. Only at the outlet of the Huskvarnaån river is there a spit built out from the west. The width of the bay is restricted to the west by boulders and blocks of the eastern end of Rosenlunds bankar. The perpendicular to the central part of the shore coincides with the maximum fetch. The difference in orientation in relation to the Jönköping bay can be explained by the sheltering effect of the eastern mainland (cf. Pl. 1). The entire bay can be fitted into a circle with a radius of 1.2 km, but the central part is too flat to fit this circle.

The shore of the Bankeryd area has developed in thick glacifluvial sand deposits on top of clayey moraine. Laterally the area can be regarded as homogeneous with respect to erosional resistance. A predominant drift towards the south-east is indicated. The perpendicular deviates 10° to the east of the maximum fetch. The shore can be said to form half of a flat bay with an approximate radius of 3.5 km.

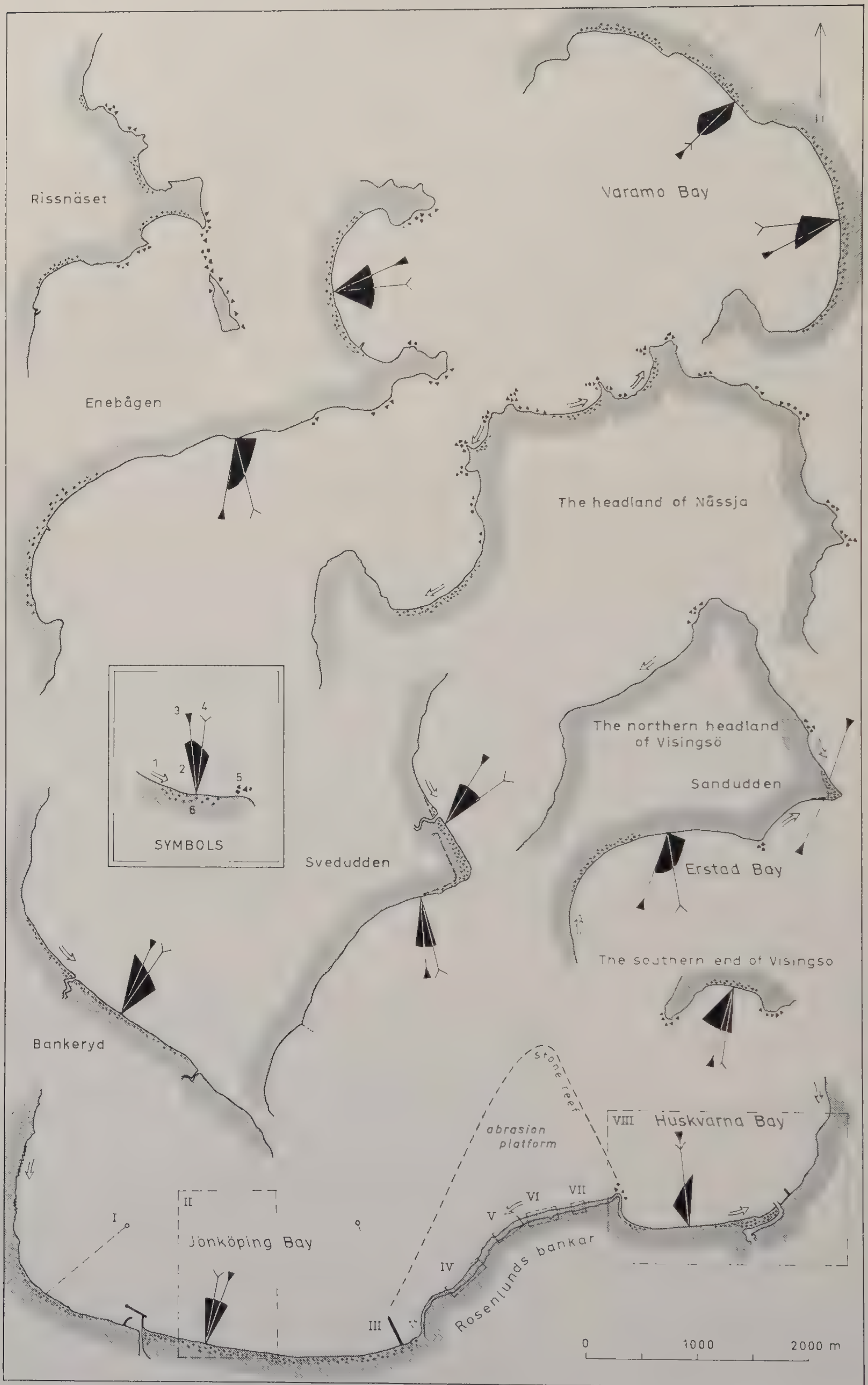
The southern end of Visingsö consists of a bay intrenched between two heads, which are rendered more resistant by outcrops of sandstone. The bay-head beach has an orientation in which the perpendicular deviates 4° to the east of the maximum fetch and 13° to the east of the axis of symmetry of the potential wave heights. The radius of curvature of the beach is 1.2 km and its width 21°.

The Erstad bay on the northern headland of Visingsö also faces southwards. It is of the half-bay type. The northern shore is cut in clayey moraine, and at the eastern end "hung up" on boulders and blocks. There is almost no beach material along this part of the bay. The north-westerly curved part consists of a beach ridge, which embays a lagoon. The exposure with regard to fetch is very skewed. The longest fetch is along the island and, of course, not effective. The central part of the northern retreating shore runs N. 81° E., and the perpendicular deviates 28° from the longest fetch. The radius of curvature is 1.6 km.

The area called Enebågen lies north of the Flisen shoal. Towards the south there is an open bay. The eastern part of this is a low abrasional shore cut in moraine with a rather high content of coarse material (cf. Fig. 75) and the shoreline undulates between more and less resistant portions. The main direction is N. 76° E. The western, nicely curved part is fringed by a sand beach that has a radius of curvature of 1.4 km. To the south it is sheltered by a small head of moraine. The longest shore-to-shore fetch of this area is directed along the western mainland shore S. 10° W. The limit of the interval of long fetches to the east is difficult to specify. The western inner part is not only exposed to waves from the south-east but also to waves from the Motala bay crossing the lake.

The position of this open bay corresponds to the position of the half-bay of Bankeryd. The latter has an inclination with the main direction of the western mainland shore of 104°. At Enebågen it is 123°. The Erstad bay has a similar position. There, the angle between the northern part of the bay and the straight shore nearest to the south is 99°, and in relation to the shore all the way down to the Visingsö harbour it is 118°. It may be noted that where the interval of long fetches is narrower and parallel to the adjacent shore–Bankeryd and Erstad–the angle is less than at Enebågen.

The outer part of the Enebågen area consists of two projecting headlands. Between these there is a deeply indented bay. The shore consists of a beach ridge with eolic deposits on top of it. Behind the ridge is a mossy swamp. The sand beach has a radius of curvature of 9.9 km and a width of 125°; this means that the centre of curvature lies in the gap between the heads. The perpendicular is directed N. 85° E., and the maximum fetch is found at


Rissnäset
Varamo Bay
Enebågen
The headland of Nässja
3
4
1
2
5
6
SYMBOLS
The northern headland of Visingsö
Sandudden
Svedudden
Erstad Bay
The southern end of Visingso
Bankeryd
stone reef
abrasion platform
VIII
Huskvarna Bay
I
II
Jönköping Bay
VI
VII
V
IV
III
Rosenlunds bankar
0
1000
2000 m

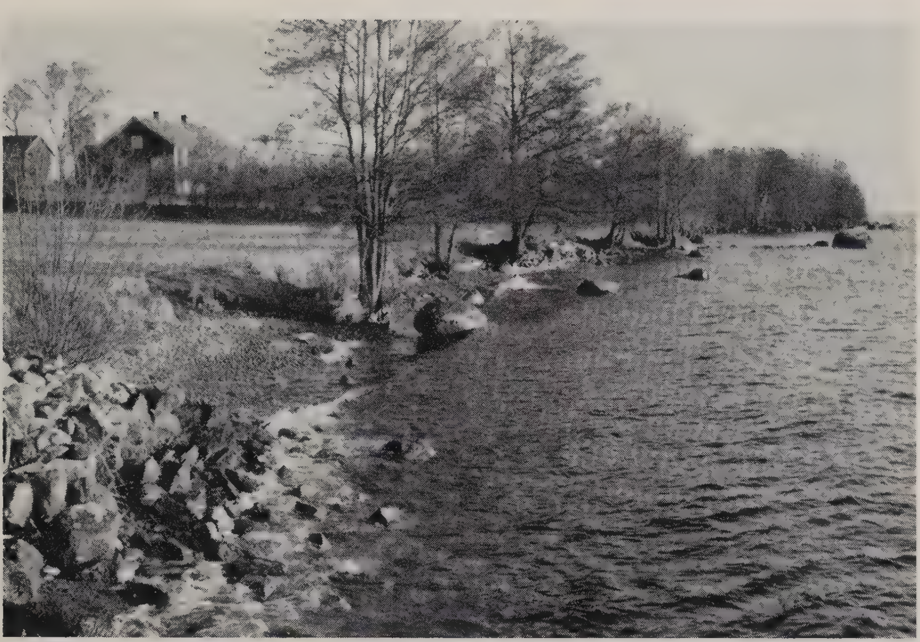


Fig. 75. Low moraine shore of the southeastern part of Enebågen. Abrasional pocket beaches with pebbles. The angular pieces in the foreground have been dumped.

N. 67° E. The deviation between these cannot be explained by the wind distribution (cf. Fig. 18, Karlsborg).

The inner part of the Motala bay is fringed with secondary bays situated between projecting heads (cf. Pl. 1), and in these bays curved bay-head beaches are found. In Fig. 74 the configuration of the north-eastern part, called the Varamo bay, is illustrated. The shore consists of three arc-shaped beaches separated by shores resistant to erosion. From north-west to south-east the orientation of the beaches turns from WNW.–ESE. to N.–S., which reflects the change of exposure. The central part of the shore has an orientation with the perpendicular directed S. 45° W., which coincides with the maximum fetch obliquely crossing the lake. The northern beach has a radius of curvature and width of 0.9 km and 57°, and the southern 1.5 km and 67° (when the part sheltered by the projecting head is excluded), and the central beach is very flat.

Fig. 74. Examples of shoreline configuration in Lake Vättern. Symbols: (1) predominant direction of drift, (2) interval of large fetch, (3) direction of maximum fetch length, (4) direction perpendicular to the shoreline, (5) boulders and blocks, (6) sand beach. Localization of special areas: I, the western section of Jönköping bay with the site of a core; II, the central part of the bay (cf. Fig. 62); III, the pier section; IV—VII, special maps of the Rosenlunds bankar; IV (Fig. 120); V (Fig. 121); VI (Fig. 122); VII (Fig. 123); VIII, map of the Huskvarna bay (Fig. 147).

Abrasional nesses

In Fig. 74 there are two examples of a projecting form with the same feature. They are the Svedudden promontory (called a promontory because it is high), which consists of sand (cf. Fig. 10), and the Sandudden ness, which is low and consists of clayey moraine (cf. Fig. 2). The shorelines of both are developed by abrasion, but there are also beach depositions, the ones attached to the northern shores being the larger. The shorelines of Svedudden have the directions of S. 34° E. and N. 73° E. The northern shore, which is almost straight, deviates 56.5° (the cliff line 42°) from the average direction of the adjacent mainland shore, and the southern shore deviates 50.5°. The perpendicular of the northern shore lies 26° to the east of the maximum fetch direction, and the perpendicular of the southern shore 13° to the east.

The northern shore of Sandudden has exactly the same direction as that of Svedudden,

S. 34° E., and the southern shore runs in its outer part N. 88° E. The cliff lines deviates 11° and 7° from the present shores. The southern shore is attached to an open bay, which in the middle part has a direction of N. 55° E. and has a radius of curvature of 0.7 km and a width of 63°.

The configuration and orientation of these two projecting forelands are most congruent. They have developed in different kinds of deposits but in positions of about the same exposure, characterized by fetches of almost opposite directions, the northern one being the longer. Their form can be regarded as an equilibrium form of retreating heads affected by two opposite main directions of wave energy. The similarity with the depositional ness with the same exposure, such as, for example, Dungeness, is striking.

Headlands

There are finally the three headlands in Fig. 74 to be described: the headland of Nässja (west of Vadstena), the small headland of Rissnäset (north of Karlsborg), and the northern headland of Visingsö. The contours of the three headlands have been drawn, in order to illustrate the configuration of retreating convex headlands with inhomogeneous conditions as regards resistance to erosion.

The northern part of the Nässja headland is made up of an ice-frontal accumulation of moraine and glacifluvial esker sediments. Some parts hold a high content of boulders and blocks, others only consist of sand and finer fractions. As can be seen from the figure, the shoreline is composed of small bays between projecting heads, and these heads are associated with the occurrence of boulders and blocks off and on the shore.

In the two bays west of the northern end the sand is found in the eastern parts, where also dunes have been formed. The distribution reflects the exposure to winds from the N.–W. quadrant. The projecting northern end is affected by winds both from the north-west and the north-east, which gives it an anvil shape. The small headland of Rissnäset also exemplifies the anvil shape of a projecting head affected by winds from opposite directions. It may be noted that the direction of the perpendicular of the northern bay is more to the east than that of the southern bay, this being sheltered to the east by a stone reef. The largest peninsula in Sweden, which has this anvil shape and consists of two opposite bays and a flat head, is the Falsterbo peninsula in Scania; there are other large peninsulas of the same form in the Danish archipelago. The typical exposure to contrary directions of high wave energy is always found.

The shores of the northern headland of Visingsö consist of a continuous series of bays and shallow indentations, orientated in relation to the exposure. The distribution of boulders and blocks of the clayey moraine is not so differentiated as in the deposits of the Nässja headland. Here the different resistance to erosion is also due to the position of the bedrock. Along the north-eastern shore the bedrock is partly exposed at the shoreline (and abraded), but even where it is not exposed, nearness to the bedrock means a higher content of cobbles and boulders. This headland and the Nässja headland are examples of the type of configuration I have proposed should be called "scalloped headlands".

Curvature of bays and indentations

The western shore of Visingsö island forms a series of very shallow indentations. In my previous study of this island (1959) the relation between the depth of the indentations and their linear breadth, expressed by the corresponding radius of curvature, was related to the breadth. It was found that the relation was significantly different on the headland and along the western shore. Linear regression showed the radius (R in metres) of the corresponding curvature to be

$$R = 1.76 + (1.39 \pm 0.16)b \tag{36}$$

on the headland and

$$R = 47.60 + (2.44 \pm 0.80)b, \tag{37}$$

on the western shore, where b is the breadth in metres. The correlation coefficients were 0.95 and 0.81. The indentations on the western shore are separated either by more resistant natural material or by artificial measures at landing places, which prohibit the retreat of the shore (Fig. 76).

The western mainland shore from the Enebågen area to the Bankeryd area is also characterized by indentations. The larger ones, which form open bays, can be seen even in the small-

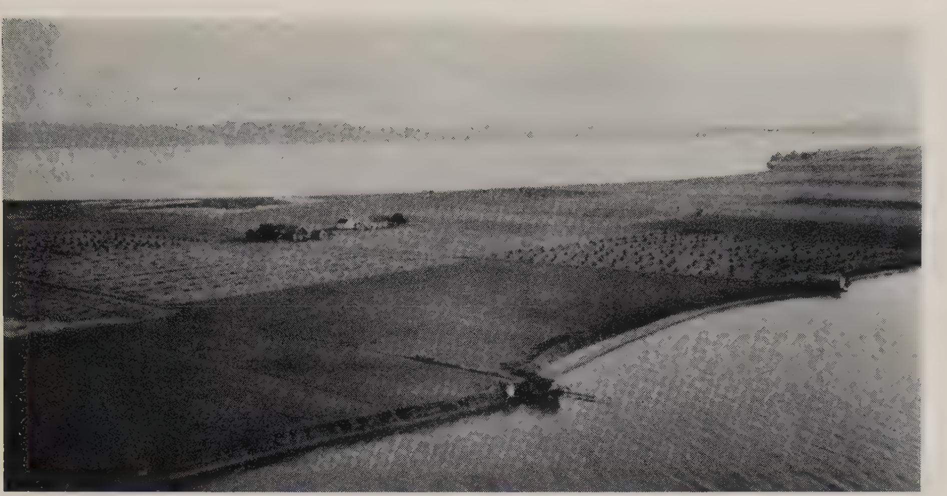


Fig. 76. Indented moraine cliffs between protected landing places on the western shore of Visingsö island. On the opposite shore the embayed lagoon of Erstad bay. Oct. 1962.

scale map of Pl. 1, especially the open bays of the Guldkroken area south of Hjo. The moraine surfaces are lower in the bays than on the headlands between them, and there are also less boulders and blocks in the bays than on the headlands. Smaller irregularities in the shoreline, not visible on the map, can also be referred to the occurrence of coarse material. The variation of the composition of the moraine possibly reflects ice-terminal positions. The same relation between moraine composition and shoreline configuration is found along the shore north of Mount Omberg. It can be stated that there is a general tendency to formation of shallow indentations along the retreating straight shores, which have a direction parallel to the long fetches.

The bays with sand shores in the areas illustrated in Fig. 74 do not form circular arcs in their whole length. Towards the ends the curvature generally increases, especially when the ends are sheltered. The curvature of shores from the map sketches, completed with some shores from the northern and southern parts of the Motala section, is illustrated in Fig. 77. It can be seen that these shores have a circular arc form, independent of size. The angular width, which fits the circular arc, is given in the legend. Ten of the sixteen shores fit the arc within a width of 50° to 70°. There are three shores with a small width (21°–38°). The smallest width is in the southern end of Visingsö, and there are three beaches with a width larger than 70°. Two of these are much intrenched and thus sheltered. The mean width of all the shores is 58.3° and, if the shore of the southern end of Visingsö is excluded, the mean is 60.7°. This implies that, on an average, the linear breadth is the same as the radius of the circular arc. These shores do not have the cycloid form proposed by Bruun (cf. p. 99) as the equilibrium form. However, all the shores cannot be fitted into a circular arc, because their central parts are too flat and thus these have a more cycloid form. Shores of this type are found in the Jönköping and Huskvarna bays and in the middle part of the Varamo bay. The half-bays at Bankeryd and Enebågen are also flat, and the same applies to the shores of the projecting heads of Sandudden and Svedudden. These bays have in common a predominant direction of beach-drifting, which also applies to the bays orientated at almost right angles to the maximum fetch. The mainly unidirectional drift means that there is little shifting of material from one end to the other, and because the wave energy is constant as far the fetch is concerned, the rate of abrasion or transportation and deposition can be expected to be much the same over the whole shore.

Conclusion

South of the mid-point of the lake the general direction of net beach-drifting is to the south. In the middle part of the western shore the

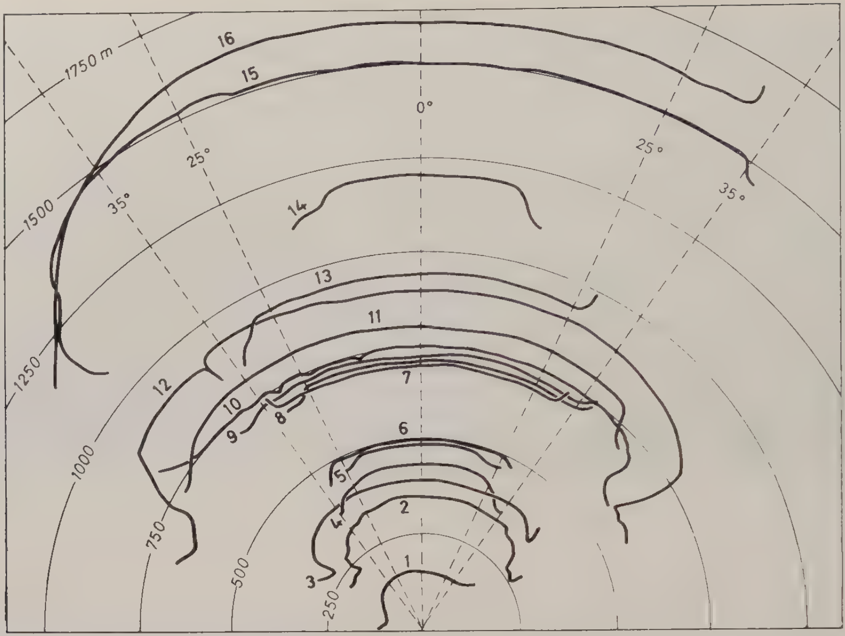


Fig. 77. Curvature of bay beaches. Shoreline contour according to the economic maps on scale 1: 10,000. The arc of circular curvature in parenthesis. 1, Hamneviken, Motala bay (70°). 2, northern shore of Rissnäset (60°). 3, southern shore of Rissnäset (75°). 4, Lilla Hals, Motala bay (36°). 5, northern end of the Nässja headland (38°). 6, Långsnapen, Motala bay (54°). 7, south of Sandudden (62°). 8, Stora Hals, Motala bay (60°). 9, south of Svedudden (70°). 10, Sandby, Motala bay (52°). 11, western shore of the Nässja headland (84°). 12, eastern bay of Enebågen (80°). 13, northern beach of Varnamo bay (64°). 14, southern end of Visingsö island (21°). 15, southern beach of Varamo bay (66°). 16, Erstad bay (50°).

direction is unspecific, and in the northern part of the lake beach-drifting is restricted by the dissected shore contour. From south of the Gränna harbour over Visingsö island to south of the Svedudden promontory there is a zone of wave-energy deficit in relation to adjacent areas. On the mainlands this zone is characterized by unspecific beach-drifting and near-shore accumulations of sand, and on Visingsö a low intensity of erosion and beach accumulations is associated with the zone.

An abrasional series of shoreline configurations has been identified. It involves: (*a*) the *scalloped bay*, made up of small secondary bays between projecting heads of greater resistance; (*b*) the *indented shoreline* of flat, open bays along mainly straight shores; (*c*) the *half-bay* developed at large steps in the main trend of the shore and orientated near to the right angle of the maximum fetch; (*d*) the *abrasional ness* of a triangular form with the two shores affected by waves in contrary directions and orientated near to half the right angle of the maximum fetch, i.e. at about the angle of maximum beach-drifting; (*e*) the *anvil-shaped head*, consisting of two intrenched bays in contra-position and a flat head; and finally (*f*) the *scalloped headland*, formed by a series of shallow indentations between more resistant portions on a mainly convex form.

Most of the bays of the lake have a circular arc form within an angular width of 50°–70°, but in some bays the central part is very flat, and this also applies to the shores of the half-bays and the abrasional nesses. The bays are orientated near to the right angle of the maximum fetch or in symmetry with the distribution of wave heights. A deviation indicates the direction of net drift.

Specific depositional features of longshore beach-drifting are only found in the southern part of the lake. The barriers or spits built out over the entrance of submerged ravines, river valleys and bays are the most characteristic.

Abrasion

Introduction

The term "abrasion" is used in somewhat different senses (cf. American Geological Institute 1960, Stamp 1961). Geikie (1898) defines it as "the operation of wearing away by aqueous or glacial action", which, applied to the shores, means erosion by waves, currents and mobile ice. Associated weathering processes are also generally included, when the abrasion of shores is described. In the chapter on shore processes in general there is no special section on abrasion, because, as regards wave action, it is a function of the processes discussed; the effect of ice is a very special effect and is dealt with in the section on shore processes in winter in this chapter.

The abrasional morphology, including the effects of weathering, varies considerably with the material abraded, the type of weathering (depending on climatic and chemical conditions), the periodic change of water stage, the general trend of water-level change, and the amount and kind of loose particles. The discussion on the morphology is therefore restricted to conditions of special importance to the Vättern shores.

The retreating shores of Lake Vättern are connected to areas of loose deposits and sedimentary rocks. As regards mechanical properties, the loose deposits can be separated into cohesive soils and frictional soils, and, as regards their genesis, into moraine, glacifluvial bed-load sediments and glacifluvial suspended-load sediments. The latter are cohesive and the bed-load sediments mainly frictional. In the moraine there is no sharp limit between the two categories, but the clayey moraine, which is most common, has cohesive qualities. When dry, it can stand in vertical and even overhanging walls, and it is so hard that it has to be worked with a pick-axe. The coarse particles of the clayey moraine make an important difference to the cohesive glacifluvial sediments with regard to the abrasion process and the resulting morphology.

According to King (1959, p. 293), solid rocks are eroded by the sea by four processes: corrosion, corrasion, attrition and hydraulic action. In Lake Vättern and in other areas with frost in winter the action of frost and ice should be added. These agencies are not included by King, who (op. cit., p. 291) gives an incomplete and thereby erroneous version of Nansen's opinion on frost action. Nansen (1922, p. 40) clearly states that, "The raised beaches of Norway have obviously, as a rule, been formed by shore erosion by frost, in a similar manner as the strandflat, but during relatively short periods". Corrosion is the same as chemical weathering, and corrasion is the wearing action of particles moving over the abrasion surface. The special form of corrasion effected by rotating particles, which form pot-holes, is often called evorsion, especially in fluvial morphology. Attrition is the diminishing of particles by breaking up and wear, and hydraulic action refers to the effect of the dynamic pressure of waves breaking directly on cliffs. King (op. cit.) states that the latter process is "probably the most effective, particularly in well-jointed rocks".

The processes mentioned also apply to loose cohesive deposits, which can be looked upon as poorly consolidated rocks. Similarities between forms sculptured by corrasion in cohesive sediments and in bedrock under fluvial conditions have been demonstrated (Arnborg 1958, p. 12). An illustrative comparison was also thought to be possible as regards the Vättern shores, but examination of the abrasion platforms in sedimentary rocks has proved that their morphology is not characterized by corrasion, and they are discussed in the section on shore processes in winter.

Most of the abrasion in Lake Vättern is in cohesive sediments, and because of the different characters of the morphology the abrasion of cohesive glacifluvial sediments and cohesive moraine is discussed separately. In the frictional soils it has not been possible to distinguish any specific abrasion morphology, and the character of the shores in areas of frictional soils is discussed in the chapter on the main shore types.

Chemical weathering

Traces of chemical weathering are common in the rocks. The rocky granite shores have surfaces granulated by more resistant crystals, especially quartz crystals. By comparison with surfaces covered with moraine it can be estimated that the Post-Glacial weathering is only at most a few millimetres.

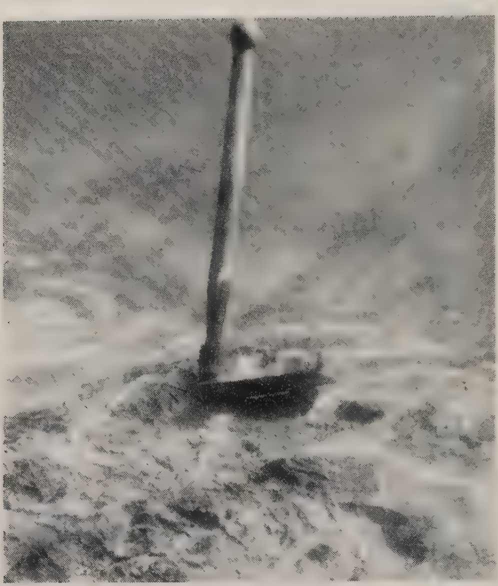


Fig. 78. Chemical weathering of carbonatic layer with quartz veins north of Visingsö island. Depth 1.5 m. Length of handle 0.5 m. July 1957.

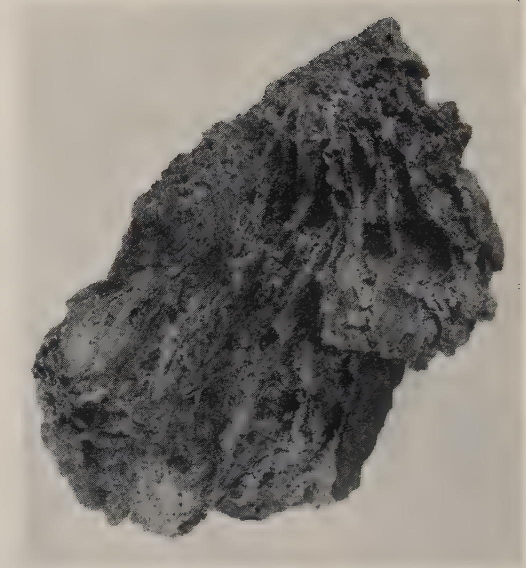


Fig. 79. Part of weathered pelitic residual block found at a depth of 8 m north of Visingsö island. 40 % of natural size.

In carbonatic strata of the Visingsö Series shallow cavities have been found (Fig. 78). On the northern shore of Visingsö such strata are exposed at the shoreline and further off. Near the shoreline, where the moraine cover has been abraded very recently, there are no cavities. The surface of Fig. 78 was photographed about 30 m from the shoreline, and it can be estimated to have been exposed for about 200 years.

There are scattered residual pelitic boulders and blocks on the flat abrasion pavement north of Visingsö. They have been observed at depths of 8 to 15 m. By chemical weathering only a skeleton remains (Fig. 79). The same type of selective weathering is found in vertical walls of outcrops examined at depths of 2 to 5 m. Even though the effect of chemical weathering is a normal feature, its importance to the development of the shore morphology must be regarded as insignificant.

Abrasion of suspended-load sediments

General

Consolidated sediments with enough clay to give a plastic quality are very resistant to erosion by water flow. The large difference between the high critical erosion velocity and the low velocity necessary for transport of the eroded fine particles (the Hjulström effect; van Straaten 1954, p. 69) is well known, both from fluvial and coastal environments.

From his studies of the bottom morphology of the river Ångermanälven Arnborg (1959 *b*, p. 6) states that corrasion is the most important erosional process of cohesive sediments. By pure shearing stress, erosion of flakes or lumps of the sediment is generally more likely than single particle erosion, because, as a rule, there are joints or bedding planes of less resistance than that displayed by the forces of cohesion between single particles. On the upper part of the abrasion platform and on the shore scarp of river banks, as well as on sea shores, where the surface is alternately dried out and wetted, breaking up by dry cracks is most important, and the loosened pieces are rounded into mud balls by attrition in the swash (Bell 1940, Morgan 1961, Nossin 1961, Norrman 1963 *b*).

Arnborg (1957) was able to study the corrasion activity directly by diving operations and has described the characteristic morphology of narrow grooves, up to some metres deep, corraded by sand in bed-load transportation, and pot-holes eroded by rotating material.

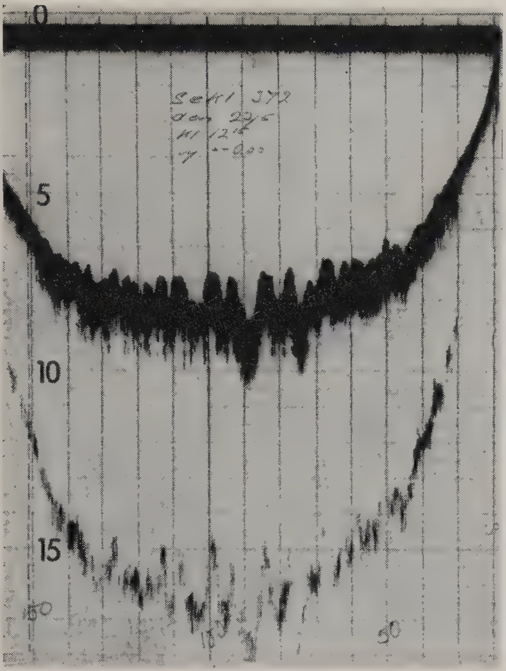


Fig. 80. Echogram of section from the River Göta Älv with furrows corraded in a cohesive sediment (from Sundborg and Norrman 1963.)

Grooves obviously eroded by corrasion of Late-Quaternary sediments on the bottom of the river Göta älv were later described by Sundborg and Norrman (1963). An echo-sounded section from this river with a corrasional ridge and groove morphology is reproduced in Fig. 80. Upstream of this section the river is shown as having cut down into frictional soil (moraine), and further upstream, where there is almost no bed load, there are no grooves. An important condition in rivers eroded in cohesive sediments is the possibility of the absence of a graded bottom profile, because there is no material that can fill up eroded hollows. In the river Göta Älv there are maximum depths, which lie 20 m below that of a downstream threshold. This condition is, of course, also fundamental as regards the abrasion of shores. By the abrasion of frictional deposits a large volume of particles that can only be transported in the nearshore zone is produced, but the abrasion of cohesive sediments means a net loss, as the fine particles will only settle permanently in deep water offshore.

The resistance to corrasion can be expected to increase with the cohesiveness, but quantitatively little is known about this relation, which can be regarded as being mainly dependent on particle-size distribution, pore space, particle shape and particle orientation. The corrasive effect also depends on the size and shape of the particles in motion and their transport velocity. The thickness of the layer of transported material determines the probability of moving particles being in contact with the abrasional surface.

The attrition of sand grains to certain dimensions has been used as a measure of the minimum diameter of particles effective in corrasion. Twenhofel (1945) considers this dimension to be 0.5 mm. This brings us to the much-discussed problem of the bimodality of sediment distributions on account of a deficit of material in the range of 2–4 mm and its relation to dynamic or mechanical instability. (A conclusive discussion is to be found in Sundborg 1956, pp. 191–194.) As regards the corrasion of the bed, it seems irrational to expect a finite limit within the grain-size interval of bed load, but the particles larger than 0.5 mm in the interval 1–6 mm can be expected to be the ones which have the greatest corrasive effect in relation to the wave energy. "It is these grains that are most easily dislodged when transport of bed load commences, and

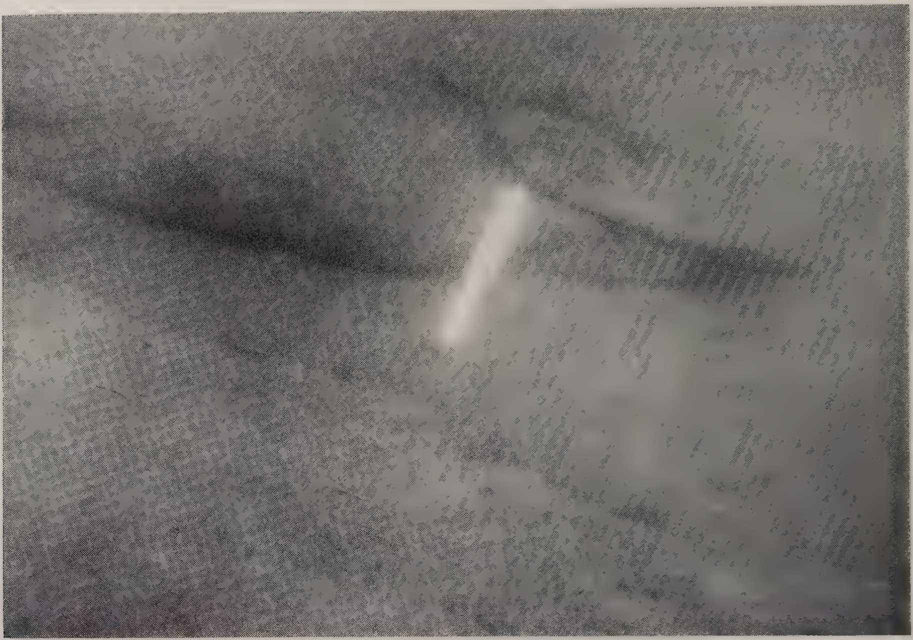


Fig. 81. Shallow groove partly filled with sand. The shore to the left. Depth 1.7 m. Length of knife 0.3 m. July 1962.

they are the last to come to rest when the movement ceases" (Sundborg 1956, p. 193).

The depth of disturbance by the wave motion determines the critical thickness of the corrasive/protective effect of a layer of sand. By experiments on natural beaches King (1959, pp. 172–176) found that with sand of a diameter of 0.22–0.29 mm the depth of disturbance increased by about 1 cm for 30 cm of increase in wave height at the break-point. On one beach with sand 0.4 mm in diameter the increase was about 2.5 cm for 30 cm of increase in wave height. From the results of these investigations it can be estimated that a 5 cm layer of medium sand has generally a protective effect in Lake Vättern, but a layer of coarse sand of the same thickness has no such effect.

The abrasion off Rosenlunds bankar

In Lake Vättern there was a special opportunity to study the abrasion morphology of cohesive sediments along the shore of Rosenlunds bankar and on the platform off this shore. A brief outline of this Glacial formation is given in the introductory chapter on the Quaternary geology,and the morphology of the present cliffs is discussed in the chapter on recent retreat. Here, only some comments on the general conditions of the abrasion will be given. The present shore scarp is cut i cohesive sediments, except for the eastern projecting end of the cliffs (cf. Figs. 11 and 74), where there is moraine. Cohesive Glacial sediments are also found on the bottom of the broad triangular platform off the shore. However, most of the bottom farther off than 60 to 150 m is covered with boulders and blocks. This residual material has not been found in the underlying sediment, which in places is exposed between the boulders. There is generally a cover of sand and fine gravel along the shore, but in periods of hard weather it is eroded. The deposit is thickest above the shoreline; on an average 0.2 to 0.3 m. Off the step it is generally less than 0.1 m, and then it gradually thins out to a distance of 20 to 30 m from the shoreline. The material of the outer part consists of sand (cf. Table 15). Off this zone there is very little loose material but a naked surface of the original Glacial sediment and scattered boulders and blocks. This sediment contains 30–50 % clay and very little sand (cf. analysis 108/61, Table 26).

The beach material is supplied from the upper parts of the cliff by the retreat of the shore. It is not moved shoreward of the breaker

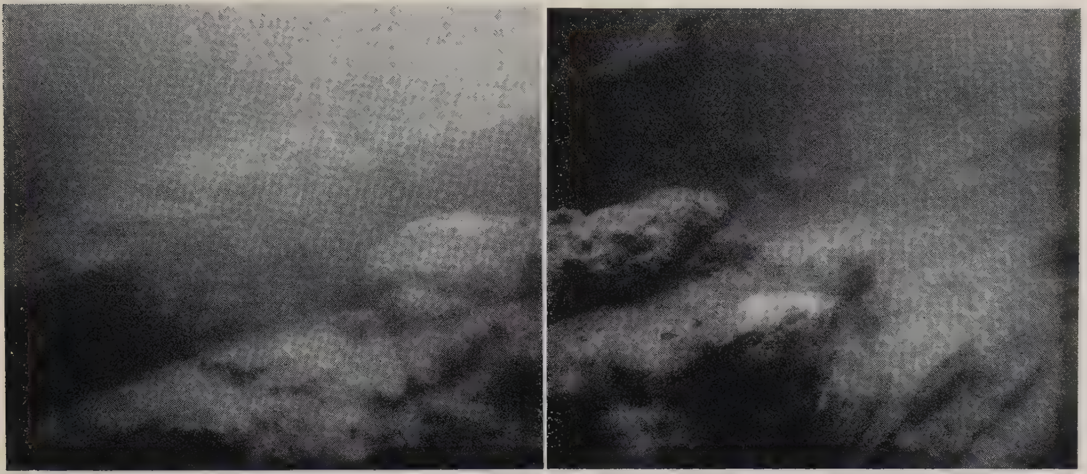


Fig. 82. Ridges and grooves of the area mapped in Pl. 3. The southern part of the two western ridges of sections 2 *a* and 3. View towards south-east. June 1962.

Table 15. Characteristics of sand in the nearshore zone off the Rosenlund cliffs. Sections 1 and 2 in the eastern part, section 4 in the central part, and sections 5 and 6 in the western part. The numbers refer to analyses of 1961. Characteristics in phi units.

Sect.	Nos.	Depth (m)	Q_1	Md	Q_3	% < 3 Ø
1	32	0.55	0.65	1.35	1.70	2
1	21	0.7	2.25	2.55	2.85	16
1	15	1.7	0.95	1.35	1.85	3
1	34	2.3	1.15	1.80	2.50	14
2	11	0.6	1.30	1.60	1.90	2
2	20	1.1	1.85	2.25	2.70	13
2	18	1.65	0.60	1.25	1.90	5
4	68[1]	0.65	2.40	4.10	6.80	66
4	33[1]	1.25	1.75	2.45	3.40	33
4	24[1]	2.3	1.40	2.30	2.90	22
5	13	1.25	0.40	0.80	1.25	1.5
5	10	1.45	0.75	1.00	1.55	0.5
5	16	1.8	0.75	1.10	1.50	0.5
6	17	2.1	0.70	0.95	1.20	1
6	12	6.0	0.60	0.90	1.25	1

[1] Material washed out from alluvial fan.

zone, but is transported along the shore into the Jönköping bay. The zone of exposed or almost exposed cohesive sediments from the outer part of the breaker zone to the area covered by residual material has a specific morphology characterized by the action of corrasion. Corrasion can also be regarded as being the fundamental abrasional process in the breaker and swash zones.

The abrasional surface in the swash and inner breaker zone is quite even. There are some shallow cavities, where the sediment is less resistant, but no regular pattern can be found. In these zones the corrading material moves both at right angles to the shore and along the shore, and thus there is grinding in various directions.

In the outer part of the zone covered with sand there are shallow flat-bottomed grooves and embryonic pot-holes. The grooves run in a direction fairly perpendicular to the shore, and the sand is gathered into them (Fig. 81). Further off they are deeper.

Sections parallel to the shore have been on p. 15. The sections are shown in Pl. 2 and map of the western third of this area has been drawn. The method of surveying is described on p. 00. The sections are shown in Pl. 2 and the map in Pl. 3. The main sections lie 5 m apart, the short sections 2.5 m from them.

In the area surveyed the grooves are corraded in a surface (cf. Fig. 83), which in the section nearest to the shore (Pl. 2, section 1) lies at a depth of 2.3–2.5 m. As can be found from the outer sections, the surface slopes very gently from the shore. The grooves have steep walls (there is no vertical exaggeration of the profiles), which can also be overhanging. Especially in sections 1 and 1 a we can see that in the broad shallow grooves deep narrow "canyons" have been cut. The grooves do not become continuously deeper off the shore, but the average level of the bottom of the grooves to the right in section 1 lies at a depth of 3.0

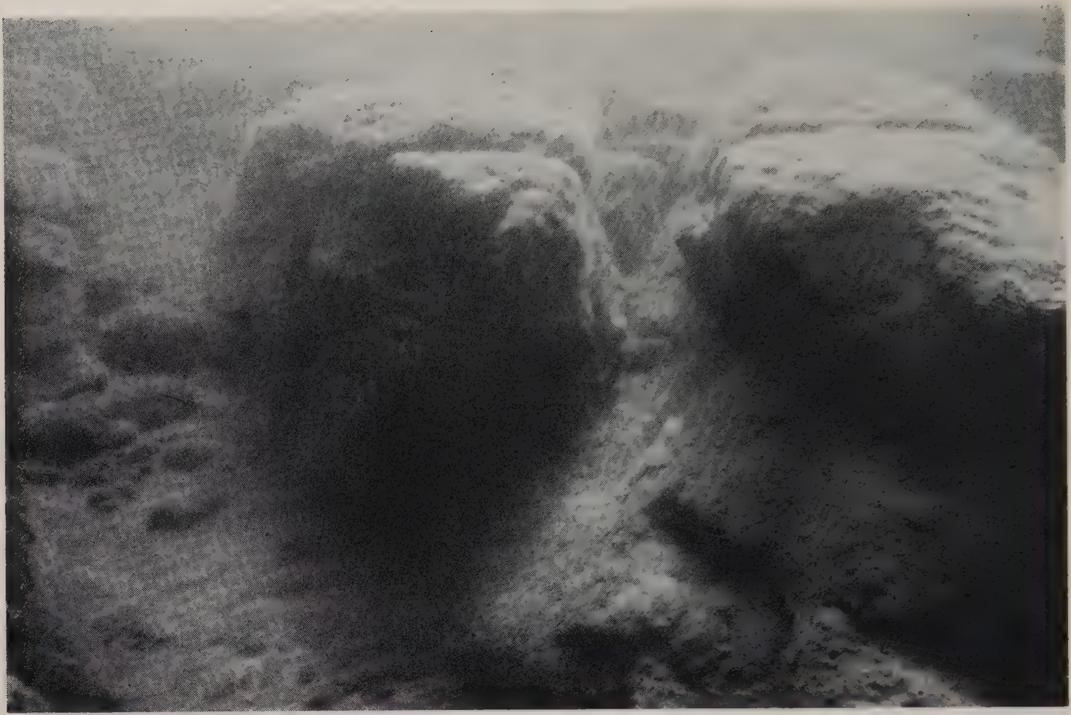


Fig. 83. Corrasion morphology in clay off the eastern part of the Rosenlund cliffs. View off the shore. Depth of upper clay surface 4.0 m. Depths of grooves about 1.0 m. Residual gramite boulders in the grooves. The clay surfaces are covered with algae. July 1961.

m and in section 6, 25 m further off, it lies at a depth of 3.8 m. In the three outer sections there is little left of the surface in which the grooves are cut. It may be noted how in areas with a cover of residual boulders the formation of grooves is prevented.

In the map (Pl. 3) the morphological pattern of the grooves is illustrated. The grooves are fairly parallel and orientated north-south, which is the direction of the maximum fetch and the maximum wave heights. The map shows that the grooves are cut in a fairly even surface. (The relief is best seen, when the map is looked at along the diagonal from the lower right-hand corner.) The abrupt end of some residual ridges, for example, between sections 1 a and 2, is connected to vertical joints in the sediment, which run perpendicular to the grooves (cf. Fig. 82). No joint system parallel to the grooves was found.

The surfaces of grooves and remnants of ridges (such remnants are found in the right-hand part of sections 2–2 a and 5–6) are much more irregular in details than could be shown in the map. At the ends of ridges there is sometimes a group of pot-holes (Fig. 84). In some of them pebbles are found, but they are often empty.

That the morphology described is the result of corrasion by sand moved by the wave motion is obvious, and it may be noted that it is much the same as the morphology developed under fluvial conditions. As soon as shallow grooves have been formed, the sand is collected in them, and the corrasion is concentrated. The areas between the grooves are little affected, because the lack of sand prevents corrasion. The walls of the narrow channels generally seem to be stable, but some pieces that had fallen down were observed.

It may be questioned if the well-developed grooves found in a zone about 50 to 150 m from the shore and at depths between 2 and 5 m are still being actively abraded, or if this is a fossil morphology preserved from periods with a lower water stage, when the shoreline also was closer. When observed during a summer of fine weather, the morphology may look

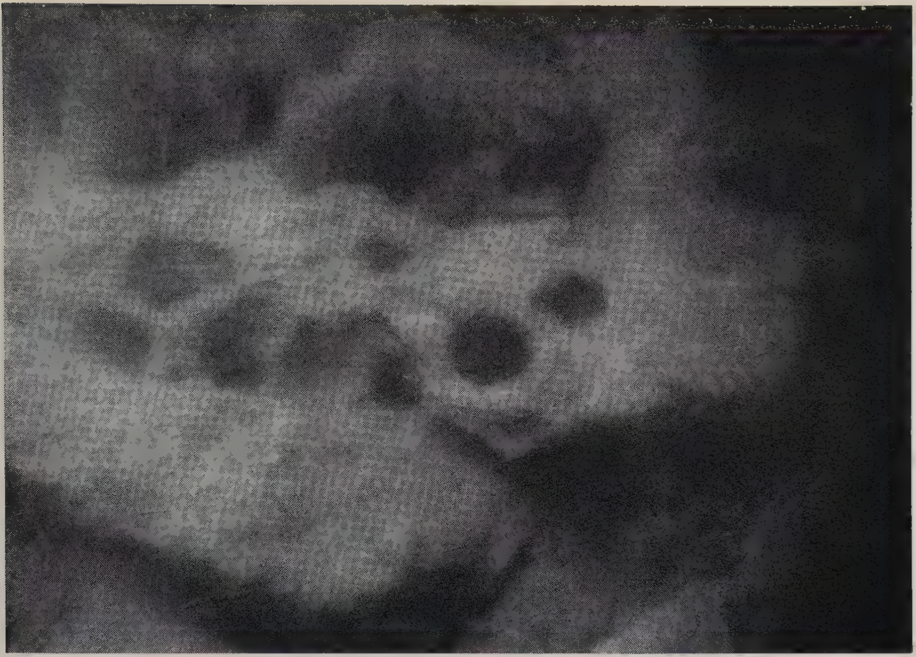


Fig. 84. Pot-holes in clay ridge east of the mapped area off the Rosenlund cliffs. Depth 2.5 m. July 1962.

very "fossil". There is little or no sand in the grooves, and all surfaces are covered with a downy coat of algae and fine particles. However, in the first storm this coat is removed, and fresh surfaces of the cohesive sediment are exposed in the grooves. The effect of the organic coat is uncertain. It has not been specially studied, but it seems probable that it has some weathering effect, and when it is removed, a thin layer of the sediment may be lost. The recent formation of this bottom morphology is proved by the map comparisons discussed in the chapter on recent shore retreat. As can be seen in Fig. 122, the area surveyed in detail was close to the cliff line at the beginning of the 18th century, and at that time the water level was only about 0.25 to 0.35 m lower than at the present time. The distribution of bed-load sediments in the Jönköping bay also confirms that the corrasion process is active. It can be stated that the grooves are developed by corrasion induced by wave motion off the zone of frequent longshore movements, which here take place down to about 1.5–2.0 m. Because it is impossible to reconstruct the original upper surface of the cohesive sediment, the total amount abraded cannot be estimated.

Abrasion of clayey moraine

The areas of abraded clayey moraine are characterized by a shallow, low, nearshore slope covered with residual pebbles, cobbles and boulders. The platform of the beach is generally higher than a line drawn in extension of the nearshore slope, and thereby the profile is somewhat concave (cf. Fig. 55, A–C), but, where there has been little or no beach material, this bench is not accentuated (Fig. 55, D). The profiles cannot be compared with equilibrium profiles, in which loose material can be shifted and adjusted to the wave action. The form of the profiles depends on the relation between the rate of shore retreat and vertical erosion, and this relation is not only dependent on the wave-energy distribution but also on the amount of residual material, the slope of the land and the rate of submergence. The slope of the nearshore bottom will increase with the rate of submergence and the resistance to abrasion. With a constant shore

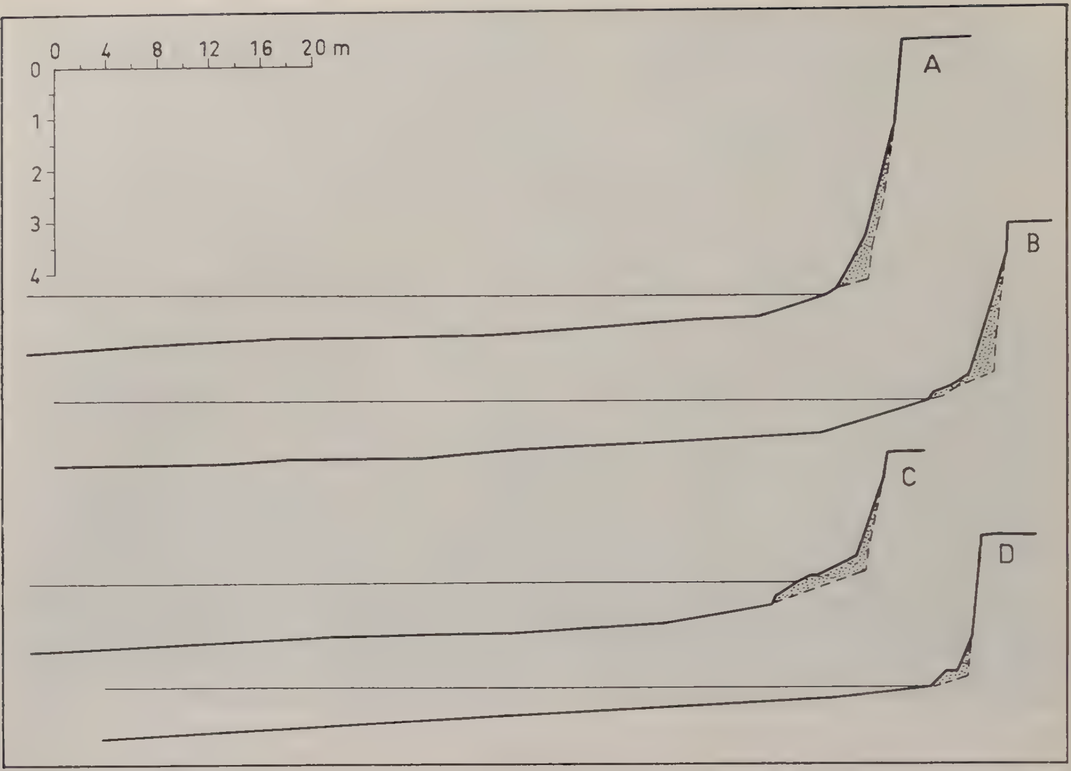


Fig. 85. Abrasion profiles from the north-western shore of Visingsö island. Talus and beach material dotted. The profiles are developed in clayey moraine, and the bottom is covered with an abrasion pavement of cobbles and pebbles.

retreat and otherwise uniform conditions the bottom profile off the concave part nearest to the shore can be expected to be straight.

In the reproduced abrasional profiles from the northern part of Visingsö the slope is extremely low (1:80, 1:100, 1:90, 1:70), and in twenty profiles from this area the mean height of the shore scarp at the cliff base was 0.7 m above the extension of the profile off the concave part nearest to the shoreline. The low slope is due to a rapid shore retreat in a flat lowland, and the shallowness of the concavity is due to a residual pavement, which makes the vertical erosion slow. Along the south-western shore of Visingsö, where the shore has been stable for a long time, the slope of the bottom mainly varies between 1:30 and 1:40 offshore from the concave part, and the height difference betwen the extension of this slope and the shore scarp is about 1.2 m. These gradients of slope are of the order normally found, but where there is much residual material, the slope may be as steep as 1:10. This is the case in parts of the moraine cliffs along the Ostrogothian plain, where the platform is narrower and the percentage of residual material higher than in the Visingsö, Gränna and Guldkroken areas.

Most of the submerged areas of central Lake Vättern have a bottom, which consists of an abrasional pavement resting on clayey moraine. The largest coherent areas are found north and west of Visingsö (cf. Fig. 126) and between the Flisen shoal and the western mainland shore. Diving operations have shown the character of the bottom to be very homogeneous, irrespective of the distance to the present shoreline. There are minor irregularities and undulations on the bottom, but in its entirety it is very even. Where there are cobbles and small boulders, the abrasion pavement is generally not thicker than a single layer, and there are patches of exposed clayey moraine (Fig. 86). Where there are pebbles, the cover may be thicker; the thickest pavements examined were about 0.3 m.

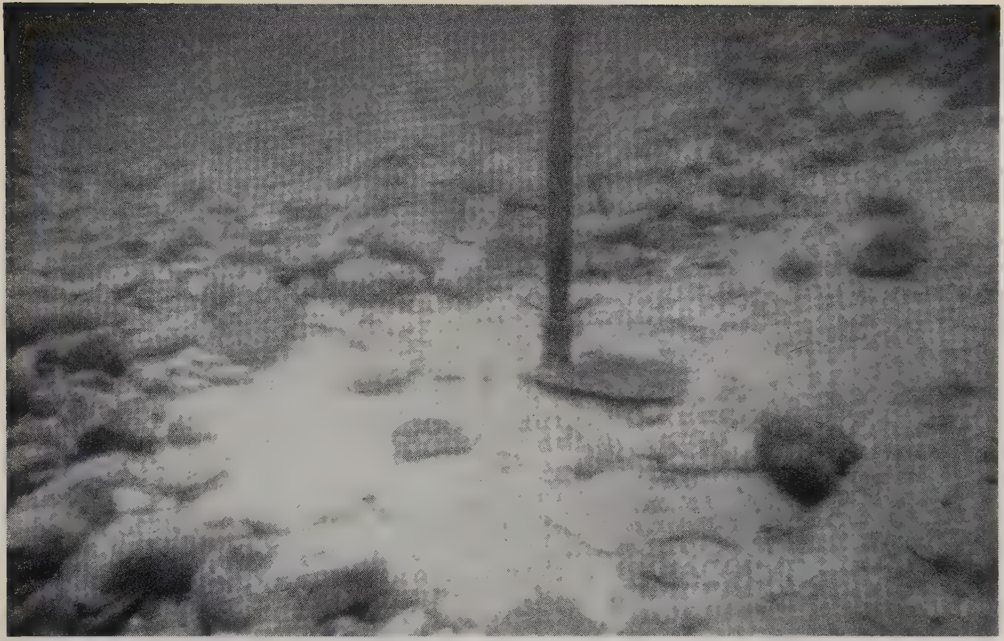


Fig. 86. Abrasion pavement of cobbles and pebbles on clayey moraine NNW. of the northern end of Visingsö island. Depth 4.5 m. Aug. 1957.

The moraine surface under the pavement and in the exposed parts is rough and hard. In some areas the plastic deformation tectonics of the moraine are prepared by selective erosion. The most distinct forms were found north of Sandudden on Visingsö at depths of 8 to 11 m and on the nearshore platform (Fig. 87) of the Vilhelmsro area on the western mainland shore, 3 km north of the Jönköping bay. In both areas there are traces of an ice oscillation.

Nowhere in Lake Vättern has a groove and ridge morphology been found in moraine. However, in Karrebaeksminde Bay in southern Zealand (Denmark) I have examined grooves and ridges formed in clayey moraine. They were found in exposed patches of moraine at depths of 2.5 to 3 m, and around these naked areas, about 10 to 30 m^2 in size, the bottom was covered with 0.2 to 0.5 m of pebbles and cobbles. The seaward limit of moraine surfaces always consisted of a steep slope of pebbles and cobbles. It seems probable that these scars in the cover were caused by heavy breakers. The ridges and grooves ran fairly parallel with the perpendicular to the shore, and the maximum difference in height between crest and bottom was about 0.2 m. There were some small pot-holes (5 to 10 cm in diameter) in the ridges.

The initial process in the formation of abrasional pavements on morainic cliff shores is clear. By undercutting, moraine falls down on the bench, this material is soaked by water, the finer particles are sorted out by the swash, and the residual material is left. If the rock is of a kind which is fairly easily attrited, the pieces must be so large that they are little moved. On the flat shores in Lake Vättern pieces of sandstone and shale have been found to keep their angular form when larger than about 15–20 cm.

Even where, immediately off the swash zone, there is a covering, resistant abrasion pavement, the surface is gradually lowered. I have examined such pavements under breaking waves and found that, although the cover is not laterally moved, the pieces wriggle, and the water becomes more and more turbid on account of the suspended material pulsating out of the cover. If the top layer of cobbles is lifted away, a thin layer, mainly of gravel, is found on top of the moraine. It cannot yet be proved with certainty, but it seems most probable that this is a local residuum produced from

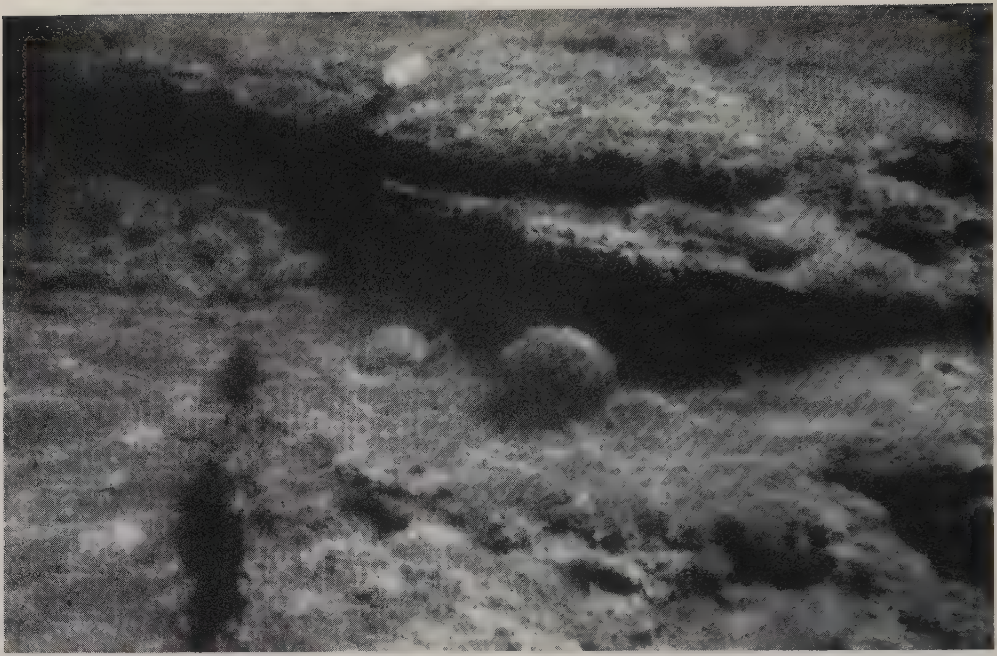


Fig. 87. Corrasion morphology in clayey moraine off the Vilhelmsro area. Depth 1.5 m. View towards the shore. July 1962.

the moraine during the lowering of the surface. The reason why this material does not penetrate the cover can be estimated to be low stream velocities under the cover and the fact that the process resembles shaking. In a shaking or vibrating process the coarser particles will stay on top.

A special proof of subaquatic lowering is found along the north-eastern shore of the northern headland of Visingsö. In this area the bedrock consists of shale, which is only exposed in some minor areas off the shore, but the moraine cover is thin and rich in the local shale. On the shore this local shale forms a small percentage of the rock species present in relation to its percentual part in the moraine. The reason is that it is rapidly disintegrated into flaky granules when dried up. However, in the abrasion pavement of the nearshore bottom there is much shale, and the pieces are strong and resistant, but, if they are taken from the bottom and exposed to the sun for a day, most of them will fall apart. Thus it is most probable that the shale pieces found on the bottom have never been exposed on the shore, but that they have been abraded from the moraine in the breaker zone.

Conclusion

By the corrasion of cohesive sediments a morphology of grooves and ridges is formed in the outer breaker zone and the inner part of the offshore zone. The abraded grooves run in a direction parallel to the propagation of the predominant waves. This morphology is of the same type as that formed under fluvial conditions. In the area studied the variable directions of movement of the corrading material in the inner breaker zone and the swash zone seem to prevent the formation of a regular pattern of grooves. The efficiency of the corrasion is demonstrated by the difference in height between the bottom of the grooves and the surfaces between them, where there is no bed-load transport of sand.

The abrasion morphology of cohesive moraine is characterized by an abrasion pavement of residual material washed out of the moraine and by low even slopes off the inner breaker zone. In the inner breaker zone the slope generally rises to a shore bench, which is the fundament of the beach. Abrasion pavements of cobbles consist, as a rule, only of a single layer, which rests on a thin layer of gravel. Pavements of pebbles as thick as 0.3 m have been examined. Abrasion pavements are estimated to cover the main part of the submerged areas of central Lake Vättern.

Shore processes and shore morphology in winter

Introduction

With respect to shore processes, the winter in subarctic areas may be defined as beginning when spray from surf and swash starts to freeze on the shores. The winter ends when the ground by the shores has thawed, and the water is cleared of ice. In other climates this season may be characterized by the steep waves of the winter storms (Shepard 1948, p. 88).

The ice conditions in Lake Vättern have already been dealt with (p. 59). Here the relations between processes and morphology under these different conditions will be examined. The description summarizes observations made at different times during the winters in the period 1954–1962.

The phenomena which have been distinguished are frost weathering, glaciation from spray and swash, ice pressure and ice drift. The influence of frost and ice on the shores can be divided into different stages, coincident with the stages of ice formation. During a winter with a total ice cover, all the stages may be run through. In winters with only a narrow belt of ice along the shores, the influence of ice drift will naturally be absent, but this may also be the case if the ice melts during a period of calm weather.

Frost

Damp ground that freezes becomes as hard as concrete. Mud flow and creep activity cease, but there are debris falls in the steep cliffs formed in Glacial deposits. Aggregates are freed by frost weathering, and talus accumulations are formed at the foot of the cliffs.

Through heavy wave attacks in the autumn, the cliffs of cohesive material often stand almost vertically, with a notch at the base (Fig. 88). If further wave attack is impeded by an ice-foot or lake ice, the talus accumulation will grow during the winter. The material accumulated will gradually shelter an increasing part of the wall from weathering. As the material falling down is spread over a growing area, the growth in height of the talus demands an increasing amount of debris. At the same time, the surface which can be weathered diminishes. Thus, the time available for weathering progressively increases towards the top of the cliff, and the wall behind the talus takes on a convex shape (cf. papers on rock talus, e.g. Lehmann 1933, Rapp 1960). Periods of snow drift from suitable directions are reflected in the talus (Fig. 89).

Frost weathering alos affects shales and sandstone cliffs, the shales more easily than the sandstones, but judging from a few years of observations, rock falls on account of frost are not frequent phenomena. The process is furthered by ground-water flow. With a few exceptions, the granitic cliffs have been very little weathered at all in Post-Glacial time.

Spray and swash

When the air temperature falls below freezing-point, spray from surf and swash glaciates the nearshore ground. Crusts of ice grow on the shores and the lower parts of the cliffs, and the ice serves as a protection against wave abrasion (Fig. 90).

An ice crust first starts to grow at the upper limit of the uprush, where there is time for the water to freeze between the few large waves in the current wave spectrum. In a stage of decreasing waves, the crust broadens towards the water surface. If only a thin cover of sand is frozen, the crust may be broken up again by new wave attacks.

As frozen sand impedes percolation, and the icy surface has little roughness, the uprush reaches further than under summer conditions. Some material brought by the uprush will be frozen fast and cannot be eroded.

As in a single wave train, the waves during a period of days or weeks have an amplitude distribution with an interval of very frequent amplitudes and a tail of less frequent, larger waves. The compound wave distribution determines the distribution of the ice crust. There is a zone with optimal conditions. This lies above the zone nearly always wetted, because it is warmed too much by the water, and below the extreme limit of uprush. At the

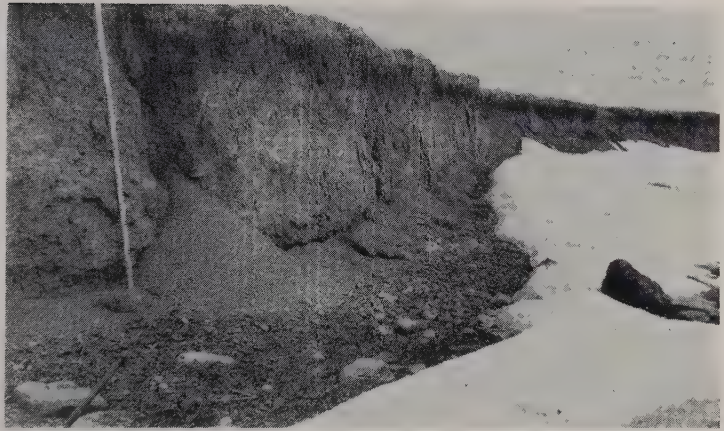


Fig. 88. Vertical cliff in clayey moraine at the northern end of Visingsö island. April 1956.

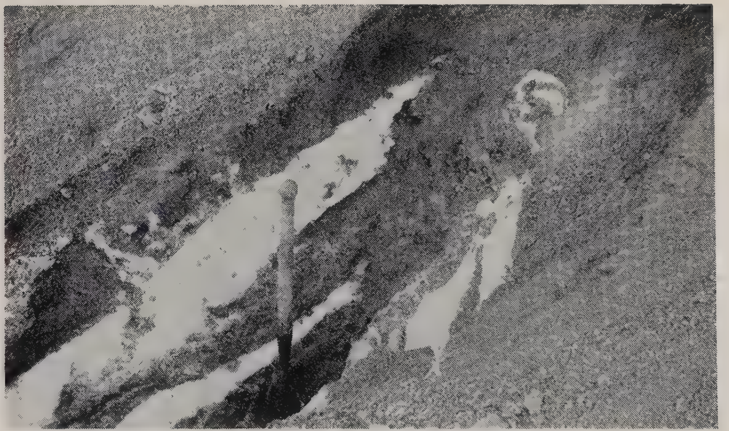


Fig. 89. Talus of clayey-moraine aggregates with intercalated snow layers. Visingsö island, April 1956.

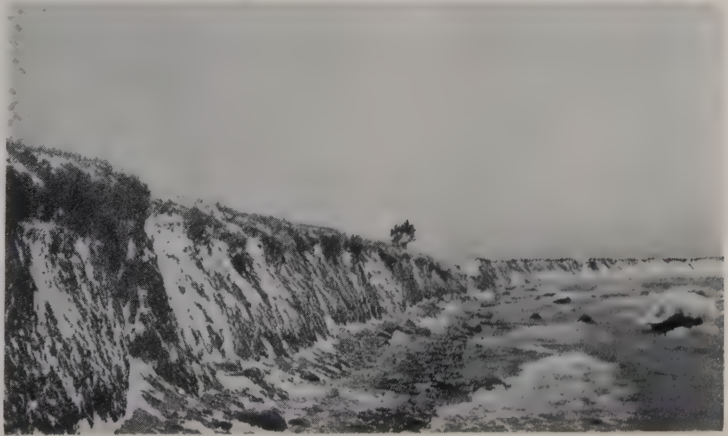


Fig. 90. Frozen cliff and iced bench on the north-eastern shore of Visingsö island. Jan. 1958.

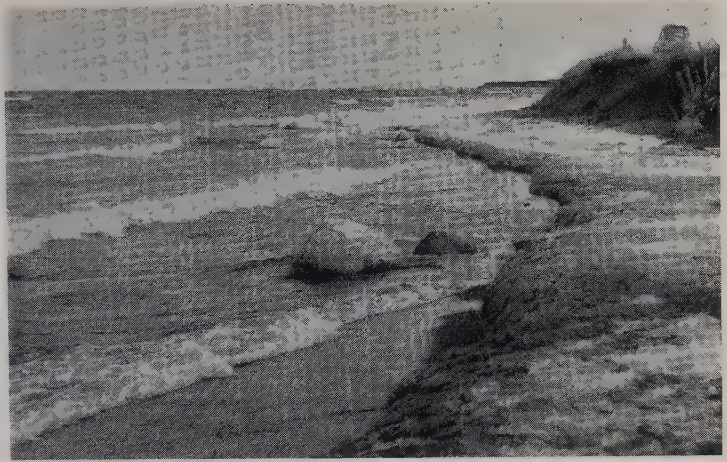


Fig. 91. Horizontal ice platform built up by swash. Western shore of Visingsö island. Jan. 1958.

Fig. 92. Ice wall built up by frazil ice and spray in the central part of Erstad bay. Note the steep front with spray gaps and the back sloping towards the shore. Arug sack on the ice. Wind 13 m/sec, temperature —14°C. Jan. 1958.

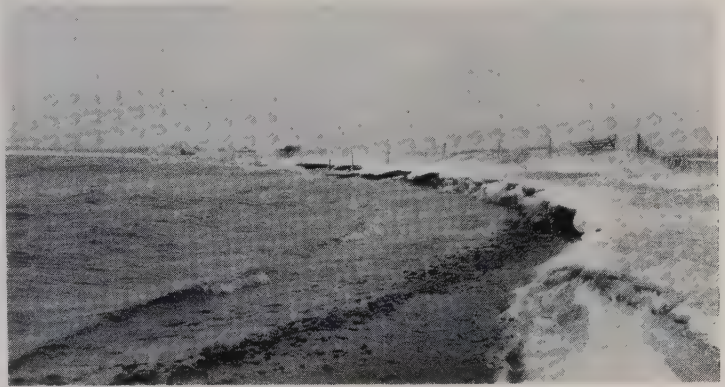


Fig. 93. Abrasion by undercutting of a frozen sand beach. Southern shore of Sandudden. Jan. 1958.

latter, the possibility of freezing is highest but the frequency of flow is too low. So the distribution of the swash makes the ice crust grow into a convex shore profile, in contrast to the normal, slightly concave profile. The crust is generally stratified by sand grains, frozen fast.

The final stage in this process is the development of an almost flat platform or bench with a steep front (Fig. 91). Its coming into existence depends on how slowly the nearshore water is frozen over.

Frazil ice in the water gives the profile a different shape. The suspension sent up the ice front by the swash is rapidly drained to form slush, and the part not washed back again is deposited at the outer edge of the ice-foot. The drained slush congeals, and a ridge of soft ice with a steep front and a more gentle slope towards the land is built up (Fig. 92).

When, in summer, beach material drifts away, the profile of the swash zone is moved landwards, keeping its characteristic morphology. In winter, when material from the lowest part of the swash zone is eroded, further remodelling of the profile is impeded by the sand being frozen, and cavities such as those in Fig. 93 are formed.

On the sand beaches in Lake Vättern, scattered stones and pebbles are sometimes found far up in unexpected positions. The most extreme case observed was on a beach ridge, embaying a lagoon in the Erstad bay on Visingsö island. Pebbles and small stones were not only found on both sides of the ridge but also far out in the lagoon on the muddy bottom. It was first thought that the stones had been thrown by playing boys, but in the following winter (1958) the true answer was found. Stones and pebbles, accumulated at the foot of the step, were washed up on the iced ridge, and some of them slid down the back and out on the ice of the lagoon, the heavier ones going very far.

This very special locality illustrates well the very low roughness of the glaciated surface and also demonstrates the difficulties and dangers of making interpretations, when the process has not been observed.

Along stony shores ice-caps may be formed upon stones projecting above the water surface, giving them a mushroom shape. The combination of stone and ice-cap has a lower specific weight than that of the stone alone, and on shores having the character of an erosional pavement, stones with ice-caps have been observed to be moved by breaking waves. This transportation is probably of small importance, but may explain otherwise inexplicable changes from one summer to another.

The effects of the glaciation of the beaches can be summed up as protective and as furthering accumulations above the water level by the embedding of material in the ice.

Ice pressure and ice drift

A wall of turfs and stones, pressed up by the ice, is a common feature around many Swedish lakes, not at least in the many small lakes in the highlands surrounding Lake Vättern. These ice-pressed walls have been described by Gustafsson (1904) and Hamberg (1919). The pressure is primarily caused by voluminal variations in the ice, due to the temperature. According to Gilbert (1890, p. 71), the process was first explained by Hitchcock (1860). Gilbert (1885, p. 109) stated that "The ice formed the ensuing winter extends only to the water margin, and by the winter's oscillation of temperature can be thrust landward only to a certain distance, determined by the size of the lake and the local climate. There is thus for each locality a definite limit beyond which the projection of the boulders cannot be carried, so that all are deposited along a common line, where they constitute a ridge or wall."

Russel, quoting Gilbert, adds the following comment (1895, p. 53): "Shore walls are not conspicuous about the margin of large lakes for the reason that they seldom freeze over and also because the winter's ice work is usually obliterated by the more active waves and currents at other seasons." This statement about the ephemeral character of the winter effects may be true as far as shingle beaches are concerned, but, while the transportation by waves is a very selective process with respect to grain size, the ice can move almost any sizes up to huge rock slabs.

In Lake Vättern ridges formed by ice pressure have only been observed in the northern archipelago. As regards the remaining part of the lake, Russel's explanation just quoted may apply. However, the fact that, south of the archipelago, the lake is not frozen over very often merely makes the lack of ridges a question of time. The true explanation is that, as long as the lake is not frozen over from shore to shore, the ice can freely expand towards open water, and when the cover is complete, the shores are covered by an ice-foot, the nearshore ice is frozen down to the bottom, and there are also ridges of drifted ice. Thus, lanes in the ice from contraction and thrusts from expansion can only be formed at the front of this protective zone.

In the chapter on the ice conditions in Lake Vättern two periods with icedrift are described, one during the freezing and the other during the melting. The conditions in the two periods have different influences on the shores. The newly frozen black ice is tough. The milky ice of the spring is brittle. On the other hand, the shores in early winter are frozen and glaciated, while in spring they are thawed. Also, the masses set adrift are generally larger in spring. The author's own observations and the experience of the local people argue that this balance favours more considerable effects upon the shores during the breaking up in spring.

Illustrative examples of boulders pressed up by ice drift are to be found on Visingsö island. Initially the boulders were left as a residuum on the shallow nearshore bottom, when the moraine was abraded. Ice drift in action was specially studied in April 1956.

In the case of a single boulder and a sheet of ice drifting ashore, the boulder must naturally project high enough to be caught by the ice. Whether the boulder will be moved depends primarily on the strength of the ice (Fig. 94). The energy of moving, packed ice-sheets, several thousands of square metres in area, is so great that a single obstacle cannot stop the drift. When the boulder is pushed towards the shore, an increasing part projects above the water level, and consequently a larger force is needed for a further movement. Thus, there is an optimal water depth for ice push, related to the thickness of the ice and the size of the boulders. Near the shore-line the ice front hits the bottom at many points and ice-floes are piled up in a ridge along the shore (Fig. 31).

Most boulders, of a mass of 0.5 to 2 m^3, are found at depths between 0.2 to 1.0 m, with about half of their mass above the water (Fig. 95). This seems to be the range of the most frequent movements. The beds of several boulders, larger than about 4 m^3, have been studied, and no indications of a movement have been found. They have stayed where they were, washed out from the moraine. So the maximum size affected by ice drift in Lake Vättern seems to be of the order of 2–4 m^3.

In some places the boulders are so regularly arranged along the shore that one gets the impression of a man-made structure that has been destroyed. Occasionally boulders can be pushed up far above the water level (Fig. 96). Farmers have told how ice has been pressed up a steep cliff, 4 m in height, and left large boulders far out in the fields.

Ridges of boulders, parallel to the shore, have also been found under the water surface in offshore position. They are relicts from times when the water level was lower in the southern parts of the lake. Hitherto, they have only been found down to a few metres in depth. It is uncertain whether this reflects changes in the ice conditions or not. Suitable areas are scattered, and it has not been possible to carry out a total inventory by diving operations. Ridges at greater depths may be hidden by sediments.

Influence of ice on bedrock cliffs

In his work on the Norwegian strand flat Nansen (1922) argued that the ice-foot (ice frozen on the shores) had a prominent erosive effect on bedrock, an effect more pronounced in fresh water than in sea water. He used the expression "shore erosion by frost" for this special form of "marine denudation" (op. cit., p. 29). In our terminology this frost action should rather be classified as weathering than erosion.

The ice-foot impedes wave abrasion, but, on the other hand, it cannot be doubted that the frost is a weathering agency with good possibilities of acting below the ice. It is as yet uncertain whether heavy wave attack on an unfrozen shore or frost weathering is in general most effective. Where no terrace is cut, or the terrace is poor in material which can corrade the surface the frost can be expected to have a comparatively larger effect.

As is described in the chapter on the rocky shores of Lake Vättern, the granitic shores

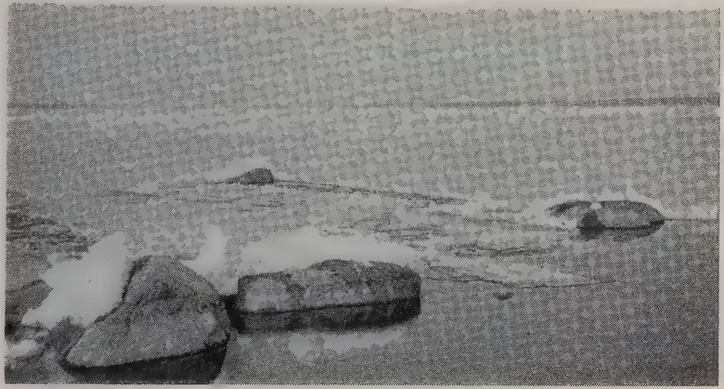


Fig. 94. Porous spring ice in drift past projecting blocks, off the northern end of Visingsö island. April 1956.

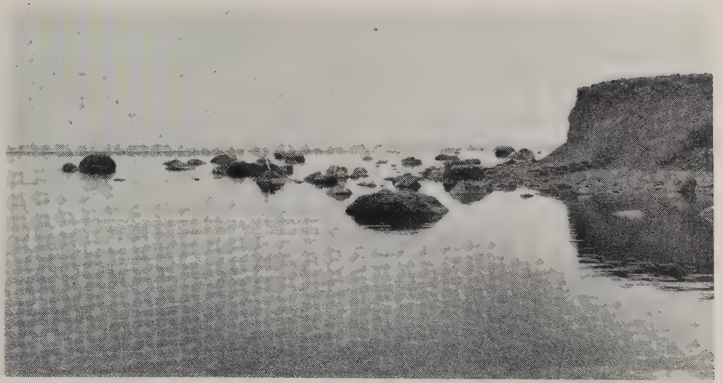


Fig. 95. Belt of ice-pushed boulders and blocks off the northern end of Visingsö island. April 1956.

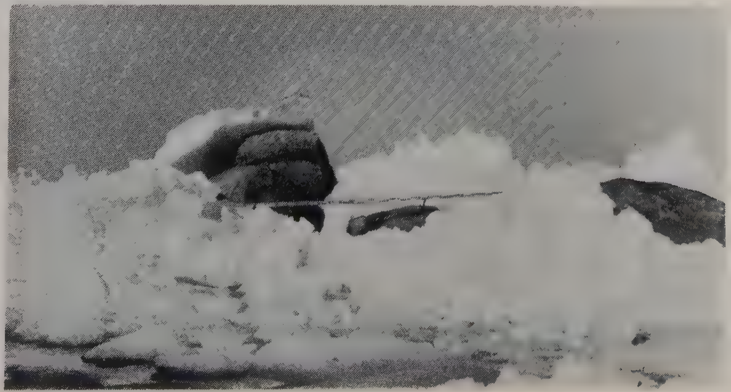


Fig. 96. Ice-pushed blocks in a ridge pressed up by wind drift on the north-eastern shore of Visingsö island. Note the old water mark of the upper block. Length of lath 4 m. April 1956.

generally have only minor marks of Post-Glacial weathering and erosion of any kind. In contrast, the erosive character of the sandstone and shale cliffs in the southern part of the lake is obvious.

Unfortunately, it is very difficult to study processes underneath an ice-foot, and it has to be estimated whether the morphology observable in summer is in conformity with one process or another.

On a rock terrace at or below the water level corrasion can be expected to give smooth forms, furrows and pot-holes, chemical weathering an irregular pattern (cf. Fig. 78) or a preparation of less resistant strata, and ice or frost weathering a breaking up along fissures and stratigraphical planes.

The sandstone and shale terraces examined have been covered either by gravel and shingle, transported by the waves, or by a layer of sharp-edged debris, or else the bedrock has been naked. Where the cover of loose material has been thick, the bedrock surface has not been examined, but where observable, it is striking, how few the corrasive marks are. The terraces are cut in planes and steps, following the strata of the sedimentary rocks, and the steps are often angular and not worn.

Along the cliffs on the south-eastern shore of the island of Visingsö, the areas of naked bedrock are especially large. However, in diving operations it was found that the bedrock is covered with blocks at depths greater than 2–3 m. These blocks, made up of the local sandstone and shale, have angular shapes. A few rounded granitic boulders were observed. Small amounts of pebbles and finer material were found in some pockets between the boulders. There were no ice-pressed ridges, and the blocks did not seem to have any definite orientation.

This bottom could be characterized as a field of block debris. It was continuously examined down to 8–9 m, where loose sediments began to hide the boulders. The material has not been transported with grinding effects. The only possible explanation seems to be that it is the product of weathering by nearshore ice, frozen down to the bottom. Pieces from rock falls may, of course, have been added, and the granitic boulders have come from the till covering the top of the cliffs (cf. Fig. 1). From the distribution it follows that the process has continued during at least the submergence of the interval corresponding to the depths examined (about 5,700 years, cf. Fig 7).

Why the bedrock is covered with more blocks at depths greater than 2–3 m is uncertain. There may have been a change in the balance between frost weathering and ice drift. A predominant drift gradually brings the material towards the shore, where it will be exposed to intensified weathering and wave abrasion.

At this locality the possibility of interference by man must also be taken into account. We do not exactly know the places where in olden times stone was quarried for building purposes, but one area known is the sandstone cliffs of south-eastern Visingsö. There was quarrying there when the castle of Visingsborg was erected in the 17th century, and most certainly also when the stronghold of Näs, situated on the south-eastern headland, was built in the 11th century. It seems most probable that loose stones and easily broken up slabs of rock were taken from the nearshore bottom. The fissure system of the rock yields excellent sizes of stones for the building of walls. Small pieces left have been worn and carried away. This "stone fishing" was probably not possible down to more than a few metres in depth.

However, the general configuration of the terrace is consistent with a continuous submergence and the present water level. Thus, the quarrying cannot have continued down to any great depth.

At the shores of emergence in northern Lake Vättern, the sandstone areas have a very different morphology (cf. p. 181) without a terrace and steep cliffs. From field examination it cannot be judged to what degree the scree of boulders and gravel is a product of frost-weathering and not only material washed out from the moraine. A smaller effect of frost weathering can be attributed to the sandstone being more compact in these areas.

A locality with a different geology, which should also be mentioned in connection with frost weathering, is the chain of islets situated at the mouth of the Motala bay. A terrace has been cut in a coarse granitic talus breccia, this morphology being most pronounced at the Jungfrun skerry (for a general description, see p. 166). In contrast to southern Lake Vättern, the water level has been nearly the same since 5700 B.C., a fact which is also reflected in the shallow flat platform. Shoal cavities in its rough, naked surface have been interpreted as scars from blocks of the breccia, the blocks having been pushed over the outer edge of the terrace by ice drift. Still, ice drift alone cannot explain the process which formed the terrace. The matrix must be weathered, and the larger pieces of rock must be broken up.

The position of the islets makes them extremely exposed to ice jam, and therefore the ice will certainly be in close contact with the bottom at a depth of 2 m. The breccia, like sandstone and slate, has weak bonds and fissures, which can be penetrated by the frost from this ice-foot of jammed ice. The combination of constant water level, exposure, and the special characteristic of the bedrock with respect to frost activity thus explains the establishment of this terrace.

Summing up, it can be stated that a morphology directly influenced by frost can be traced in the terraces cut in the talus breccia of the islets off Motala and in the shales and sandstones of southern Lake Vättern. In the latter, the main indication of this frost action by ice frozen down to the nearshore bottom is the frequent occurrence of angular boulders and large slabs, which partly fit well into the bedrock surface. The presence of a cover of blocks down to great depths is explained by the submergence. With time, the cover has an increasing possibility of not being disintegrated by weathering. Already at a depth of 1–2 m, the chance of freezing is small. Pebbles and finer material are washed out by the waves, but the coarser material will remain and be embedded at depths where fine material can be deposited.

Conclusion

Among the processes and phenomena described in this chapter, the following can be stated as the most essential.

The protective effect of the ice on beaches during the season when the winds are most violent. Without this protection, long sections of the shores would have the appearance of a steep cliff, ascending directly from an abrasion platform, and the rate of erosion would be much faster.

The transportation of boulders towards the shores by wind-generated ice drift in spring. The fringe of boulders, pressed up along many shores, is a striking feature in the landscape, and, because of the size of the boulders, could not have been created without the force of the ice.

The frost weathering of cliffs in cohesive, glacial deposits and of terraces, cut in breccia, sandstone and shale. Compact, consolidated, cohesive soils are very resistant to erosion and abrasion, but, when the material is broken into pieces, it is rapidly disintegrated by soaking and wear. The effect of the ice frozen fast to the rock terraces is not impressive in the same way, because, when the process is going on, it is invisible, and the retreat of these cliffs is so much slower. However, considering the broad terrace, often marked predominantly by this kind of weathering, and the very short time of creation (from the geological point of view), it can be questioned if this is not the most important morphological feature due to the climate.

Coast and shore classification

General

In a lake, even though of an appreciable size, the dimensions of the marginal morphological features are limited, and the zone in which the morphology is directly determined by the waves or is a direct consequence of the wave activity is, as a rule, well defined. Unfortunately, there is no generally accepted term for this morphological zone.

General classifications refer to the seaboard conditions, and are mainly based on the concepts of coast and beach. The author does not intend to review all the contradictory opinions as regards the definition of these terms, which can be found in dictionaries and handbooks (J. W. Johnson 1919, Valentin 1952, Wiegel 1953, Guilcher 1954, King 1959, A.G.I. 1960, Stamp 1961), but wishes to point out that the term "beach" is generally used in a very restrictive sense. "A beach is an accumulation of loose material around the limit of wave action" (King 1959, p. 1). As regards the width of the beach, King includes the subaquatic sediments directly affected by wave motion, but many authors restrict the seaward limit to the mean low-water line (Wiegel 1953). On the other hand, the term "coast" or "coastal area" refers to large, indefinite areas, more or less influenced by their vicinity to the sea, and this term can hardly be applied to morphological features on lakes and other inland waters.

The term "shore" is much used by some authors, whilst other authors mainly use it in compounds such as shore processes, shorelines, etc. This term is sometimes very restrictively defined as, for example, by J. W. Johnson (1919, p. 160): "... the zone over which the water line, the line of contact between land and sea, migrates; ...". But it is often used in a very wide sense: "... the strip of ground bordering any body of water" (Wiegel 1953). It is essential that the term "shore" should not include any restrictions as regards the material. The shores may thus be rocky as well as sandy.

In this paper the term "shore" is used as a general concept, and is applied to the zone above the water line, where the morphology is directly determined by the waves or is a direct consequence of the wave activity. Thus it includes shore cliffs and slopes, dependent on the shore processes, and also dunes of sand blown from beaches.

The subaquatic area affected by shore processes is divided into a nearshore zone and an offshore zone. The nearshore zone extends outwards from the shoreline (the water line) somewhat beyond the breaker zone (cf. Wiegel 1953), and is the zone, where the bottom is directly affected by a wave motion capable of initiating bed-load transportation. In the offshore zone the bottom is mainly affected by wave-generated currents developed in the nearshore zone, and all transportation, except gravity movements, takes place with the material carried in suspension.

All types of classifications of coasts, based on submergence and emergence (J. W. Johnson 1919), on the relation between terrestrial agencies and marine processes (Shepard 1937, 1948), or on advance and retreat in connection with the trend of present relative water-level movement (Valentin 1952), can also be used for shore concepts.

Shore classification in Lake Vättern

Since the ice recession the Vättern basin south of Motala has been affected by shore processes during a rapid emergence and a slower decelerating submergence. North of Motala there has only been a still-active emergence. This effect of unequal uplift has greatly influenced the present shore morphology as well as the offshore morphology in areas covered by the zone of shore processes. Shores of active retreat or shores with a morphology characterized by retreat are in the majority. There are a few areas prograded by shore deposition, and they are found south of Motala. In the part of the lake continuously influenced by emergence advance has only taken place by exposure of surfaces previously covered by water. The variation in the present shore morphology is a function of the reaction of different rocks and soils upon shore processes and water-level movements.

Although coastal classifications could be used, the classification of the shores needs a more detailed differentiation, based on erosional and depositional features related to the material affected—the source of the processes. On retreating shores the beach is an ephemeral constituent. During active recession it is lacking, and the subaerial shore morphology is characterized by the cliff or slope directly bordering upon the nearshore zone. On advancing shores the development behind the beach is mainly independent of the shore processes, but in areas where the trend has changed from retreat to advance remnants of the erosional morphology will survive for a long time, because the slope processes are generally slow.

In Table 16 the main morphological shore features found in Lake Vättern have been placed in relation to different sources and advance or retreat of the shores. The morphological characteristics in the table are expressed as concisely as possible, and should not be looked upon as complete.

The rocks have been separated into crystalline and sedimentary rocks. The crystalline rocks have been insignificantly affected by shore processes during the development of the present shores, while in sedimentary rocks a specific shore morphology has been formed.

The deposits have been separated into the two main groups of morainic and sedimentary deposits, and these two have been subdivided into frictional and cohesive soils. The moraines and the sedimentary deposits differ significantly as regards sorting and range of grain sizes. The important differences between frictional and cohesive material, as regards processes, have previously been stressed, and the differences are equally important as regards form elements. Sandy moraines contain a larger proportion of material, which can serve as a beach source, and a larger proportion of boulders and stones big enough to become a residual material than the clayey moraine, which has a cohesive character. The frictional soils cannot remain *in situ* on slopes with a steeper angle of repose than about 32°–36°, while the cohesive soils can form vertical cliffs. The frictional soils may be resistant to erosion on account of their content of coarse material, but the matrix is easily eroded. The cohesive soils become more resistant with decreasing grain size.

The sedimentary deposits differ from the moraines in their lack of boulders and stones, as regards the frictional soils, and the total absence of coarse material, as regards the cohesive soils. Very coarse material may be found in glacifluvial sediments of eskers and ice-marginal deltas, but such features are of minor importance along the present shores of Lake Vättern.

The advance and retreat have been grouped into changes connected with deposition and emergence and abrasion and submergence. The morphological shore categories distinguished by this classification are to some extent dependent on the relation between the rate of water-level movement and the initial slope of the border land. During a submergence or emergence the time available for swash action (per metre of the slope) will be less on a gentle slope than on a steep slope.

Under "retreat, connected with abrasion", are included all types of shores dominated by abrasion, even though submergence may have influenced the morphology and the rate of the retreat. If the abrasional effect is only superficial (a slight re-working of the top layer), but the main characteristics of a formerly developed morphology are unchanged, the retreat is classified as being connected with submergence.

The shores classified as marsh and meadow in areas of submergence and emergence are found in sheltered areas, where the vegetation is the most characteristic shore feature. These shores with a belt of sedge and reed have not been specially studied.

The nearshore features have been placed in the same systematic classification (Table 17) as the shore features. Phenomena caused by abrasion predominate in the table, and to a still greater extent they predominate in the areal distribution. The deposition terraces classified under retreat and advance, connected with submergence and emergence, are mainly apparent as a fringe of coarse material that has been washed down and deposited at the foot of shore slopes. The mud and sand without specified morphology correspond to the marsh and meadow shores. As regards the crystalline rocks in areas of submergence, the morphological forms of the table naturally do not denote the rock as such, but the deposition of material at the foot of rocky shores.

A comparison of the two tables of shore and nearshore morphology shows that there is a limited number of combinations. Quantitatively the constituents of each type vary, but it has been found possible to arrange the shores, the nearshore zone included, in the nine main shore types listed. They have been grouped in four categories. The cliff shores have marked abrasional features, and they are all found in the area of submergence. In the flat-shore group areas of both submergence and emergence are represented. In the former areas the shores have a depositional character, or a recession has proceeded without forming any cliff. The sandy shores with a scarp are found in areas where the changes in water level have have been small, and the shores characterized by vegetation are found in sheltered areas, where the wave activity has been of minor importance. The raised-slope shores are the shores of emergence, where the slopes are marked by shore features originally generated in the swash zone but gradually raised beyond the limit of wave activity. The rocky shores include all shores of naked bedrock which have not been significantly affected by the shore processes. The bedrock may have been primarily naked, or a cover of loose material may have been washed away by the waves.

The main shore types of Lake Vättern

Rocky shores

Cliff shores formed in
Rorainic cohesive deposits
Morainic frictional deposits
Sedimentary frictional deposits underlain by cohesive deposits

Table 16. Morphological shore features in relation to shoreline movement and sources.
0 indicates that the combination does not exist in Lake Vättern.

			Sources					
			Rocks		Morainic deposits		Sedimentary deposits	
			Crystalline	Sedimentary	Frictional soils	Cohesive soils	Frictional soils	Cohesive soils
Shoreline movement	Retreat connected with	Abrasion	0	Cliff Residual material	Steep slope (cliff) Residual material	Cliff Residual material	Steep slope (cliff) Residual material	Cliff
		Submergence	Naked rock	0	Residual material Dune ridge	Marsh (Beach ridge)	Shore scarp	Marsh
	Advance connected with	Deposition	0	0	Barrier			0
		Emergence	Naked rock	0	Residual material Stone pavement	Meadow (Residual material)	Dune (Stone pavement)	Meadow

Table 17. Morphological nearshore features in relation to shoreline movement and sources.
0 indicates that the combination does not exist in Lake Vättern.

			Sources					
			Rocks		Morainic deposits		Sedimentary deposits	
			Crystalline	Sedimentary	Frictional soils	Cohesive soils	Frictional soils	Cohesive soils
Shoreline movement	Retreat connected with	Abrasion	0	Abrasion terrace Residual material	Abrasion terrace (plane) Residual material (Subaquatic bars)	Abrasion terrace (plane) Residual material	Abrasion plane (terrace) (Residual material) Subaquatic bars	Abrasion plane (terrace)
		Submergence	Deposition terrace Subaquatic talus	0	Residual material Stone pavement	Mud Sand	Subaquatic bars	Mud
	Advance connected with	Deposition	0	0	Subaquatic bars			0
		Emergence	Deposition terrace Subaquatic talus	0	Deposition terrace	Mud	Sand Residual material	Mud

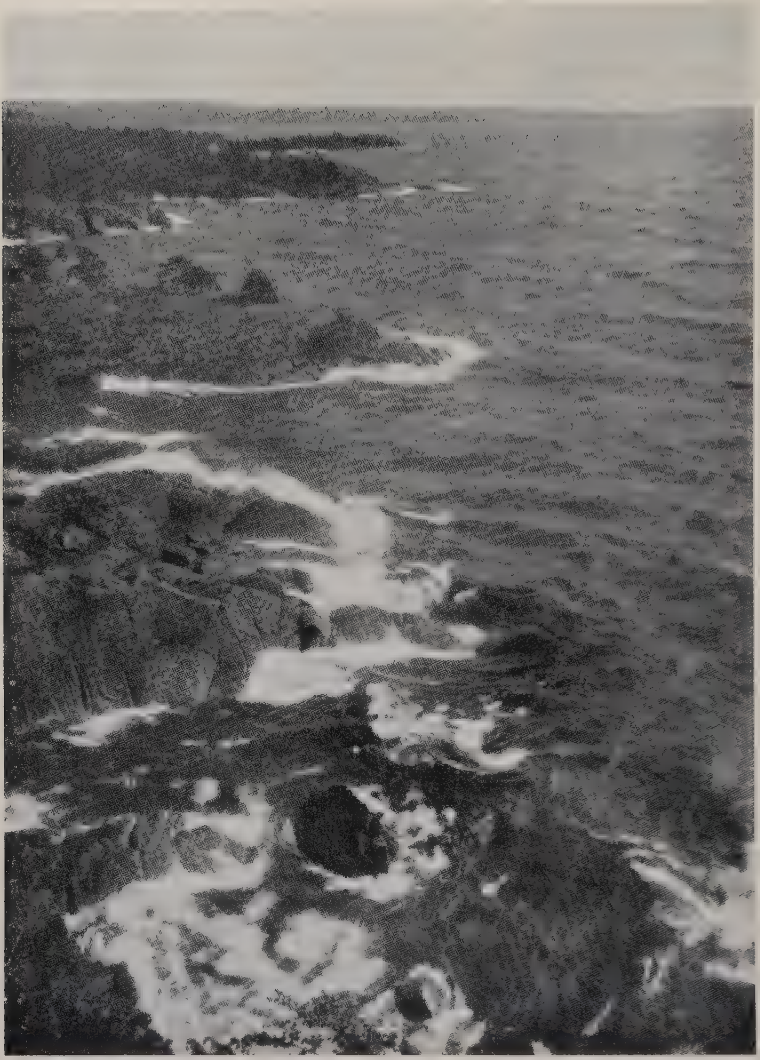


Fig. 97. Rock shore of the Harge promontory in the northern archipelago (from Stålberg 1939).

Flat shores

Shores with a dune ridge
Shores with an abrasion scarp
Morainic meadow and marsh shores
Barrier beaches

Raised-slope shores

In the following section the main shore types are described separately, and then an outline of their regional distribution is given. A description of an abraded moss is added, although this moss is not included as one the main shore types.

Description of the main shore types in Lake Vättern

Rocky shores

As regards form, the rocky shores can be divided into several groups—shores of different height, steepness, roundness, dissection, etc. These characteristics, however, are of a "pre-shore" nature. The shores are rocky, because the bedrock is exposed, and its exposure is due to steepness or wave-wash, but the bedrock, as such, is generally little or not at all influenced by the shore processes, as far as the crystalline

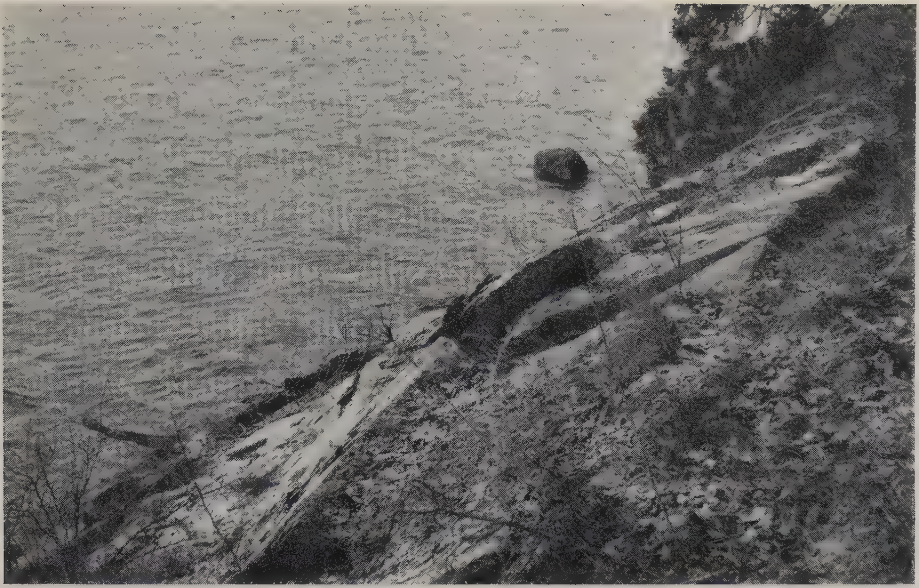


Fig. 98. Rock slide in shale on Mount Omberg. Nov. 1963.

rocks are concerned (Fig. 97). Thus a systematic differentiation of the rocky shores with regard to form is of no interest from the viewpoint of the relation between shore processes and shore morphology. An outline of the bedrock morphology was given in the introductory chapter on the geology of the Vättern basin, and further details will be found in the regional descriptions of the shore morphology. Here some typical features associated with rocky shores will be briefly discussed.

From the Swedish coasts it is well known that any appreciable abrasion of crystalline rock is only found in areas where the rock is dissected by joints and fissures (cf. Ahlmann 1916, Behrens 1953). A large part of the subaerial erosion can be ascribed to frost weathering. There is no abrasion of crystalline rocks in Lake Vättern on a scale comparable with this erosion on the sea coasts.

In steep granite cliffs with horizontal joints, notches formed by fallen blocks can be observed, but this is merely an accidental feature. In the archipelago of the northern section the shores facing south are generally more littered with blocks than those facing northwards. The former are more exposed, but they also form the leeward side in relation to the glacial motion. It has been found impossible to distinguish with certainty between blocks freed by glacial plucking and blocks freed by Post-Glacial shore processes. In my opinion, most of the blocks are products of Glacial erosion.

The only areas, where the shore processes have had a pronounced effect on crystalline rocks are in the island chain at the mouth of the Motala bay and along Mount Omberg. In the first area, already described in the section on cliff shores, the abrasion has taken place in a talus-breccia, and along Mount Omberg the rock is much fractured and brecciated. Here, seven shore caves and some smaller cavities are to be found.

The caves of Mount Omberg have been described by Munthe (1920, pp. 48–50). He says that they have been "turned out" by the action of the waves. He also mentions that the bottom of the outer part lies below the water surface and the inner part above it. Generally, the length of the caves does not exceed 9 m; in the largest one, the Rödgavel cave, 12 m of its length lies below the water surface and 5 m above it. Munthe also states that the depth at the mouth of the caves varies between about 1.5 to 2.4 m.

All the caves have cross-sections shaped like a triangle standing on its base. The size of the cross-section diminishes quite linearly. Most of the caves have a width at the mouth of 9 to 10 m, but a few are considerably narrower.

The three caves along the northern part of Mount Omberg were examined in detail in a diving operation. It was found that the bedrock at the mouths lies 1 to 1.5 m below the water surface, that half to two-thirds of the length of the caves lies above the water, and that above the water there are only minor traces of abrasional grinding. The slope along the length of the caves varies between 5 ° and 20°. The steepest slope is in the inner part. The bottom below the water level was lowered by erosion for a maximum of 0.5 m and generally not more than 0.1–0.2 m. Here the bottom had smooth hollowed forms, the cavities being partly filled with a mass of grinding stones. Above the water the bottom surface was mainly rough, and the inclined walls were entirely rough.

These caves owe their existence primarily to frost weathering of a suitably tectonized bedrock. Almost everywhere along the Omberg shore ground-water percolates along fractures, and large fractures corresponding to the inclined walls and with water oozing out are found at the end of the caves.

The grinding stones have been formed from rock pieces, which have fallen down. The slope beside the caves is too steep to permit any transportation along the shore. The wave surge has carried most of the fallen material out of the caves, and this process has been more important than the evorsion. The caves can be estimated to have been formed at water levels mainly within the range of the present water-stage variation.

In connection with this description of the Omberg caves it seems convenient to discuss the shore morphology of the Visingsö Series in the same area. The rock found along the central Omberg shore is mainly shale, but sandstone also occurs. The strike is conformable with the escarpment, and the westerly dip varies from almost vertical in the northernmost part to about 20° in the southern part (cf. Blomberg 1905, pp. 18–19). Where the slope is steepest, the shale layers form a naked wall, but most of the slope is covered with weathered rock debris.

Through abrasional undercutting the slope becomes unstable, and rock slides are released along strata of least resistance. The debris-covered slope is marked by rock-slide scars, entirely or in part covered with vegetation (Fig. 98). The slide material forms a fan of blocks, and from their orientation it is obvious that there has been movement in a body.

In the northern part the abrasion platform is only a few metres in width, but to the south it broadens to at least 15–20 m. Because of the cover of boulders and stones it is difficult to state the exact width. The bedrock has been observed down to about 2 m. It would be possible to free the bedrock surface by underwater operations. However, this would take a considerable time, because it is very laborious to carry away the material by hand, and this work can only be done during good weather.

It can be debated, whether the Omberg shore as a whole should be classified as a cliff shore or as a rocky shore. Obviously, there are parts which have a true cliff character, but there are other parts that do not have it. The same problem exists in some other areas of crystalline rock, for example, the steep rock walls south of Gränna and along some parts of the Tiveden Forest. However, the shores, which have a morphology that is more dependent on the "initial" topography and subaerial denudation than on the shore processes, have been classified conservatively as steep rocky shores.

It may finally be briefly concluded that in the southern and central section of the lake there are only steep rocky shores with the following distribution: along Mount Omberg, from Hästholmen to 10 km north of Gränna, and on the headland 9 km south of this town. Along the western shore there are outcrops on the headland south of Bankeryd and in Brandstorp at the latitude of the northern end of Visingsö. In the northern section there are only cliffs of this character in the southern part of the Tiveden Forest north of Rissnäset. The main part of the rocky shores of the northern section is not steep, but have the character illustrated in Fig. 97 with the bedrock surface marked by Glacial wear and plucking. This type of shore is predominant in the northern part of the Tiveden Forest and in the archipelago.

Cliff Shores

Sedimentary rocks

Cliffs and abrasion terraces cut in sedimentary rock are frequently used in handbooks and text-books to illustrate the power of wave

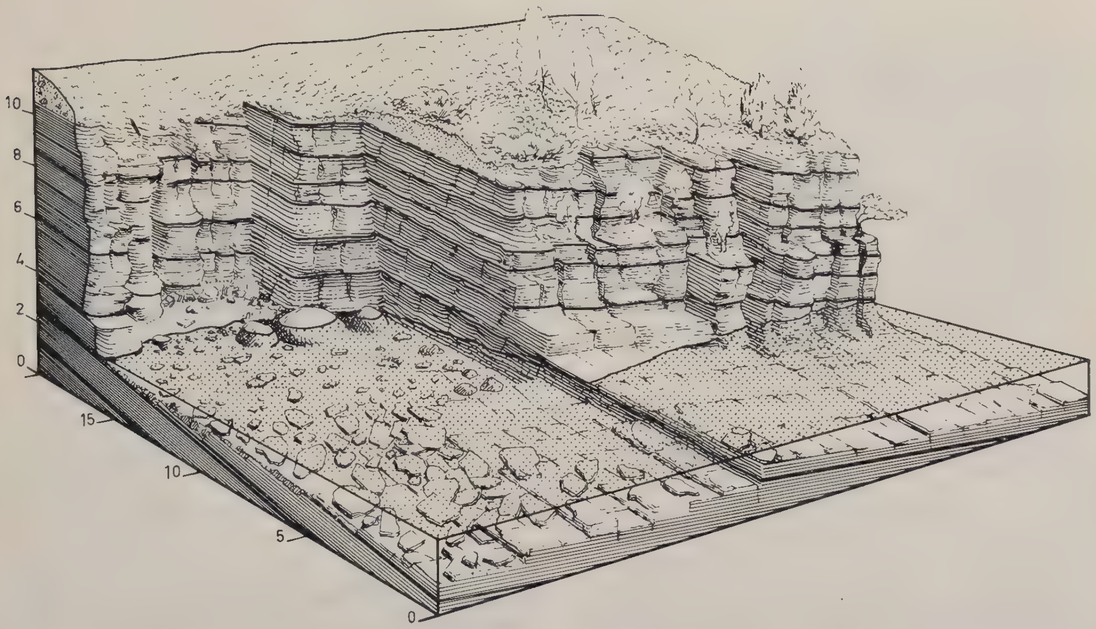


Fig. 99. Cliff abraded in sandstone. Selective abrasion along bedding planes and vertical joints. The bottom slope is lower than the dip of the strata. The diagram is based on profiles from the south-eastern shore of Visingsö island. Scales in metres.

abrasion. The difference between the resistance to abrasion of crystalline and sedimentary rocks is well marked in Lake Vättern. The morphology, illustrated in the block diagram of Fig. 99, is based on profiles and observations from the south-eastern shore of Visingsö, but the general features of the diagram are representative of all cliffs of this kind found in the southern part of the lake. The bedrock of the area illustrated consists of arkoses and fine-grained sandstone beds. The strata dip about 10° from the shore. The bedrock is dissected in a rectangular pattern by vertical joints, running N. 50° W. and N. 40° E.; the former system is most marked and makes an angle of 70° with the main shoreline direction. The joint system gives the shoreline a stepped configuration.

The cliff walls nearly parallel to the main shore direction descend in narrow steps with surfaces determined by the dip of the strata; the walls perpendicular to these follow joints and are almost vertical. The top of the cliff is covered with moraine.

The cliff foot lies slightly above or below the mean water level. The morphology of the scarp varies. It can be smoothly rounded or sharply angular, or there can be a shallow notch. In some parts the abrasion has selectively followed the joints transverse to the shoreline, and formed narrow caves, which penetrate 3 to 4 m into the cliff.

The cover of loose material at the cliff foot is thin or entirely lacking, except where there has been a recent rockfall. Scattered crystalline blocks, which have fallen down from the moraine cover, are found along the shore.

The nearshore bottom slopes about 3.5° (1:15). This slope thus has about one-third of the dip of the beds (cf. Fig. 100). The bottom partly consists of the naked bedrock, partly of broken-up slabs of rock and smaller pieces worn by abrasion. The difference between the dip of the beds and the bottom slope comes out in the morphology of the naked bedrock. The bottom follows the surface of a stratum for a few metres, then comes a descending step, facing landward, after which the bottom follows a lower stratum, then comes another step etc. The height of the steps depends on the interval between weak boundaries of the strata, and it varies from about 0.05 to 0.2 m. The distance between the steps is related to the joint system. The morphology of this abrasion surface is

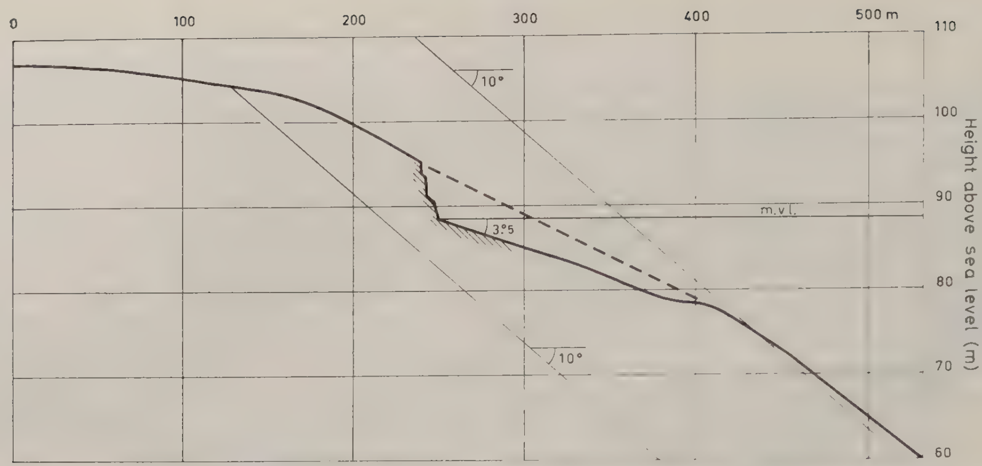


Fig. 100. Profile from the south-eastern shore of Visingsö island. Dip of sandstone beds of the cliff and nearshore bottom 10°.

discussed in the chapter on shore processes in winter, where it is stated that the abrasion is largely due to frost weathering by lake ice frozen down to the bottom. The rather steep nearshore and offshore slope reflects the slowness of the retreat as compared with the cliffs formed in loose deposits.

On the cliff shores of the mainland the amount of debris is generally large enough for the development of a narrow shingle beach. The larger amount of material can be ascribed to a thicker moraine cover, to rivulets that carry material to the shore, and to longshore drift from areas with a morainic cliff.

A different abrasional morphology is found on the islets and skerries at the entrance to the Motala bay. This could also be expected, because the bedrock is a coarse talus-breccia (cf. p. 27), and the water level has been nearly the same for about 7600 years because of the position near the outlet at Motala (cf. Fig. 7 and Norrman 1964). Three attempts have been made to map the nearshore bottom around the Jungfrun skerry (K in Pl. 1), but each time the author was forced by bad weather to abandon the work. The description is based on observations made during a diving operation in 1960.

Around the skerry there is an abrasion terrace with an asymmetrical shape in relation to the present shore. Facing west, its width from north to south is 15 to 20 m, but to the east the width is only 1 to 2 m. The depth of the broad terrace varies between 1.5 and 2 m, and at its outer edge the depth is at most 3 m. At the cliff the bottom ascends in a steep smooth curve. There is no terrace surface above the mean water level, except on the northern shore, where there is a narrow strip. This is also the only place where loose material is found on the terrace. The material consists of extremely well-rounded, granitic stones of an average size of 0.1–0.2 m. The cliff walls are steep and rugged.

The level surface of the broad, naked terrace is partly polished, partly rough. The rough parts are mainly found in shallow cavities, which have a depth of one to a few decimetres and a diameter of 0.5 to 1.5 m. These cavities can only be explained as scars from blocks of the breccia.

The outer edge of the terrace is sharp, and the subaquatic slopes steep, sometimes vertical. The slopes to the south have a morphology characteristic of glacial plucking. At depths larger than 7–8 m blocks were observed, lying as a steep talus without any cover of fine material.

The distribution of the width of the terrace coincides with the distribution of wave energy (cf. Fig. 35), and the depth of the outer parts of the terrace indicates that the water level has never, in Post-Glacial time, been much lower than at present. The abrasion process cannot be proved in detail from the information available. The distribution of rough and polished surfaces indicates that continuously the

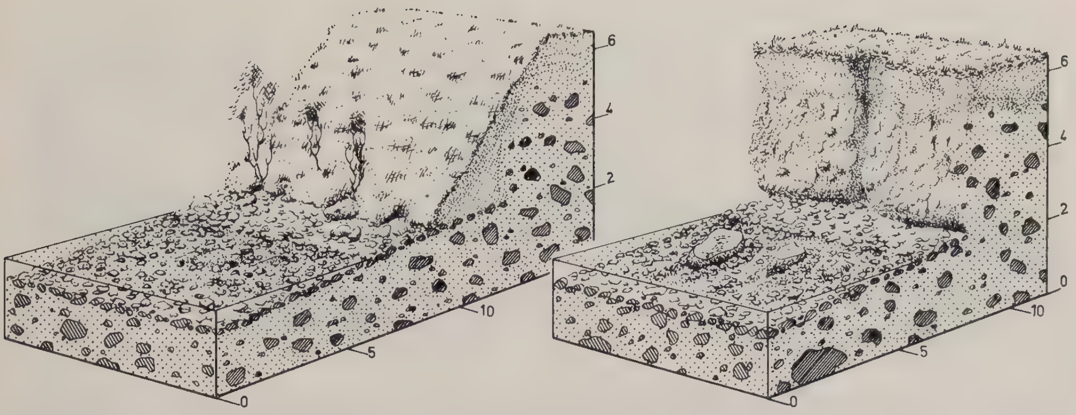


Fig. 101. The morphology of cliffs formed in cohesive moraine. To the right, in a stage of active retreat with an abrasion pavement all the way up to the notch; to the left, in a stage of temporary stability with the talus covered by grass. Scales in metres.

matrix between the blocks of the breccia has first been weathered, and then wave abrasion has polished the hollows. This is a process analogous to that found along joints of other sedimentary rock cliffs and in caves on Mount Omberg. The blocks have been detached and brought over the edge of the terrace. The possibility of loose material being left on the open terrace surface has been very small, and the corrasion there has been insignificant. Therefore the scars left by the blocks have not been polished. What force has detached the blocks? In the chapter on shore processes in winter the morphology has been interpreted as a result of ice drift, and if the blocks found on the steep offshore slope come from the terrace, no other force can explain this erosion and transportation.

Cliffs formed in the Visingsö Series are found along the south-western shore of Visingsö from its southern end and 2.7 km northwards, in the Gränna area from 2.2 to 9.7 km north of the Gränna harbour and in isolated outcrops 3 to 6 km south of the harbour.

The few outcrops of Paleozoic strata of the Ostrogothian Plain are broken up by glacial abrasion, and the transition from bedrock to moraine is vague. The cliffs have more the character of moraine cliffs than of bedrock cliffs.

Morainic cohesive deposits

This shore type has been illustrated by two pairs of block diagrams, based on profiles from the island of Visingsö (Figs. 101 and 102). The main material of this type of shore consists of a glacial deposit with a percentage of fine silt and clay, sufficient to give it a cohesive character. There may be intercalated lenses and a top layer of sedimentary clay, and the top may also be covered with a veneer or reworked, sandy material. The cohesive, glacial drift has the following characteristic qualities. It is resistant to soaking. It can stand in vertical walls, from which material is obviously detached by frost-weathering and falls owing to undercutting. Creep and flow is of secondary importance. Less than about 20–30 % of the material can be described as a beach source The amount of cobbles and boulders is hardly enough to develop a single-layer pavement.

The rock pieces of the moraine are of different kinds, and have a variable resistance to weathering and abrasion. In southern Lake Vättern the percentage of rocks from the Visingsö Series is high in the cohesive deposits, and the same applies to the content of Paleozoic rocks on the Ostrogothian Plain. When dried, fine-grained pieces from the middle and uppermost stages of the Visingsö Series are rapidly weathered, the quartz sandstones, as well as the Paleozoic limestones, are more resistant, but they are worn down by grinding. Most resistant are fine-grained granites and gneisses, but they form a smaller proportion of these deposits.

The right-hand diagram of Fig. 101 shows the morphology of a cliff during recession

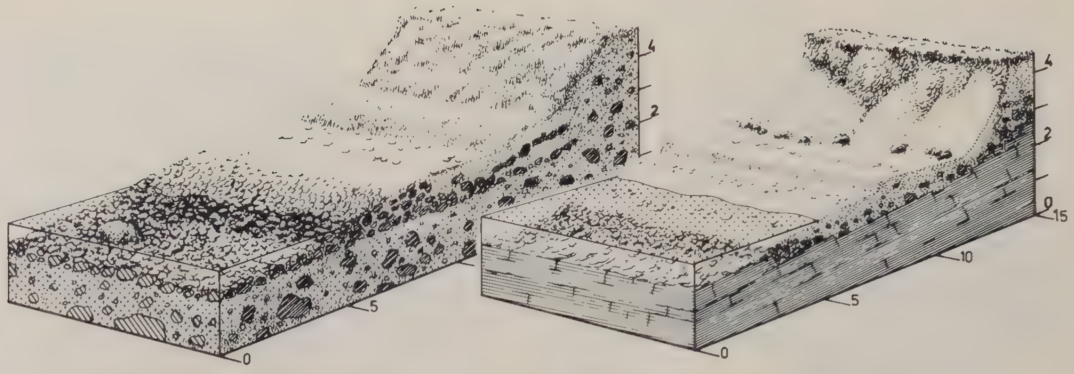


Fig. 102. The morphology of cohesive moraine cliffs with a beach. To the left, a cliff that has been stabilized for a long time by a mixed shingle and gravel beach. To the right, a cliff in which the bench is abraded in shale bedrock. During the temporary shelter of a sand beach a talus slope of weathered moraine pieces has been formed. Scales in metres.

caused by direct wave activity. The water level is 20 cm above mean w.l., and the foot of the cliff is submerged by 10 cm of water. There is no beach material, and the abrasion plane of the nearshore zone extends right up to the steep cliff. Its level illustrates the fact that the wave-cut surface is found at about 10–30 cm above the mean w.l. at the scarp.

Through undercutting a notch has been formed, and pieces of the cliff wall are falling down. The nearshore bottom has an even slope. It is covered with a protective stone pavement, and there are also some ice-pressed residual blocks. As there is no outer edge to the abrasion surface, the term "abrasion plane" has been preferred to "abrasion terrace". On shores of this type, studied at Visingsö, the slope of the offshore zone and the outer parts of the nearshore zone may be as low as 1 : 100 (cf. Fig. 85).

The left-hand diagram of Fig. 101 shows a cliff covered with talus material. This material has been weathered from the cliff wall, and owing to the residual material of the abrasion plane and less exposure to swash, the talus has been preserved. If there is time enough for the growth of grass and bushes (during some years of low water stage), the slope will be more resistant to erosion. In such a state the slope may give a false impression of shore stability, when in reality there is an abrasion of the nearshore zone off the cliff.

The two diagrams of Fig. 101 show the alternative morphology of this type of cliff shore, which because of drift has no beach or only very temporary beach accumulations. The one with the talus slope is always preceded by a period of vertical cliff, and the change in the shape will be repeated. The length of time for the alternative states corresponds to the rate of recession. A sudden rapid abrasion of a shore which during a human lifetime has been looked upon as stable has sometimes been taken as a proof of a general rapid recession or even a rapid acceleration, as, for example, in Blomberg's reports on the Visingsö shores (1908), while in reality it reflects a discontinuous retreat, made up of periods of standstill and active recession. This makes it difficult to estimate the average recession from observations only covering a short time.

The morphology of shores developed in the same material as the previous ones but with a beach which has permanently occupied the shore for a long time is illustrated in the left-hand diagram of Fig. 102. The beach consists of shingle and some sand. Counting of the shingle on the south-western shore of Visingsö (Ella Nilsson 1957) showed that for a distance of 3 km in the principal downdrift direction (southwards) the percentages of material of the Visingsö Series continuously decreased from 85 % to 25 %, and the percentages of Archaean rock increased in the same proportion. Because of exposed beds of sandstone, the percentage of Archaean rock again fall to about 60–70 % near the southern end of the island. This differentiation with regard to resistance to wear has been accentuated by the permanency of the beach, which has made possible a long period of grinding motion up and down the beach. Because of submergence and abrasion, the nearshore bottom off the step on this shore lies significantly deeper than on the more rapidly retreating shores. Carbon–14 dating has proved that the shore of this area has been stable for about 500 years (cf. p. 188). During this time the submergence has been about 65 cm. In 1957 a rapid erosion started. This erosion can be reckoned to have been caused by the combined effect of submergence and a deficit of material available for the longshore drift.

The cohesive moraine cliffs have finally been represented by a shore where the abrasion has cut down to, and even into, the underlying sedimentary bedrock, represented in the right-hand diagram of Fig. 102 by shales of the Visingsö Series. The diagram is based on profiles from the north-eastern shore of Visingsö. Besides this abrasion surface, characteristic features of the profile are the sand beach and the talus.

The mean slope of the nearshore bottom of the beach is 1: 70 (0.8°), and the dip of the strata is about 1: 10 (5°–6°) off the shore. The bedrock has been weathered by the frost action of ice frozen down to the bottom, and the detached material has been abraded. At the foot of the cliff the bedrock surface lies at the mean water level and 0.5 m above the extension of the bottom slope. On the cliff this surface lies on an average 0.6 m above the m.w.l. Thus it lies about 1 m above the extension of the nearshore bottom slope. The abrasion of bedrock in the nearshore zone cannot be said to have been 1 m, but the profile proves that there has been, and still is, an abrasion of the bedrock. The bottom is covered with residual material.

The sand beach has been deposited by beach-drifting, and it frequently shifts. Sometimes all material is removed, sometimes the beach has a width of 9 m. In this environment pieces of shale have no chance to survive, and there are few loose pieces on the bedrock underneath the sand. On the beach surface, small shale fragments (0.5–2 cm) are accumulated in bands at the upper limit of the swash. The contrast between the number of pieces of shale in the zone affected by swash and in the zones further off shows that the residual material of the latter is largely the product of frost-weathering off the swash zone.

Steep cliffs formed in cohesive moraine are found in the western part of Rosenlunds bankar, in the Vilhelmsro area on the western shore north of Jönköping, on Visingsö island, in the Gränna area, and along the Ostrogothian Plain north of Mount Omberg.

Morainic frictional deposits

This shore type owes its character to the frictional quality of the material and to the content of boulders and blocks of this material. The block diagram of Fig. 103 is a synthesis of profiles from the area north of Hjo.

The cliff slope stands at an angle of repose of about 30°, but, where material has been washed down to the cliff foot, the slope may be considerably less. The shore scarp is occupied by residual boulders and blocks, and there is a narrow beach mainly constituted by residual material. The configuration of the nearshore bottom depends on the variable amount of residual material. The diagram shows a pavement partly covered with patches of drifting sand.

Because of the amount of coarse material, these shores are generally quite stable. Ice pressure has contributed to the development of a natural apron, which is resistant to wave action. At high water fine material behind this apron may be washed out, the residual material of the beach may then be dissipated, and recession may continue, until enough residual material has again been accumulated.

The quality of the material is generally not entirely frictional; especially in the deeper parts of the deposits there is some cohesiveness, and pronounced cohesive material is often found beneath the pavement of the nearshore zone. This probably reflects differences in the composition of the moraine during different stages of the deposition. Detailed irregularities in the shoreline configuration in the form of small promontories are due to local high contents of boulders and blocks. These accumulations can be seen in the cliff, as well as in ridges and hillocks on the bottom all along the shore of the Guldkroken area. The largest block and boulder hillock examined was a shoal immediately off the harbour of Hjo. It

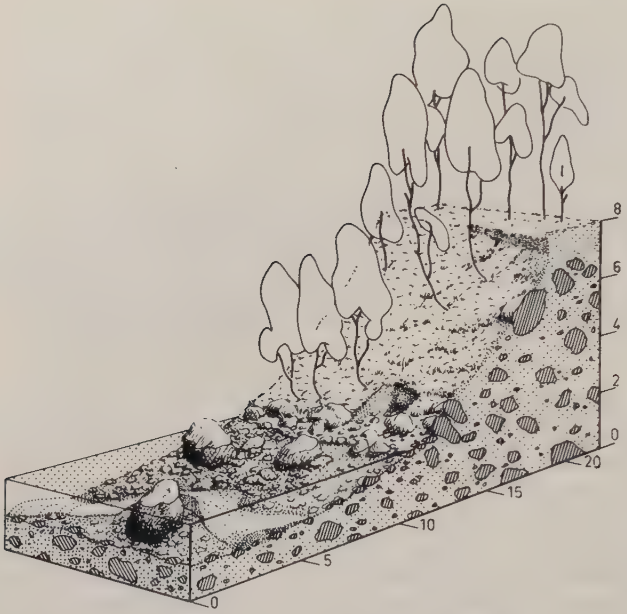


Fig. 103. Cliff formed in frictional moraine. Residual boulders along the shore and a small erosion scarp at the shoreline. Creep movements in the slope. Scales in metres.

has a height of 4 m and a diameter of about 150 m.

The described type of moraine cliff is found from about the southern end of the lake to about the latitude of Motala in the following localities: south of the Bankeryd area, along Guldkroken, on the peninsula of Karlsborg and in the bay north of that, on the Nässja headland (Fig. 104), and along the shore north of Huskvarna.

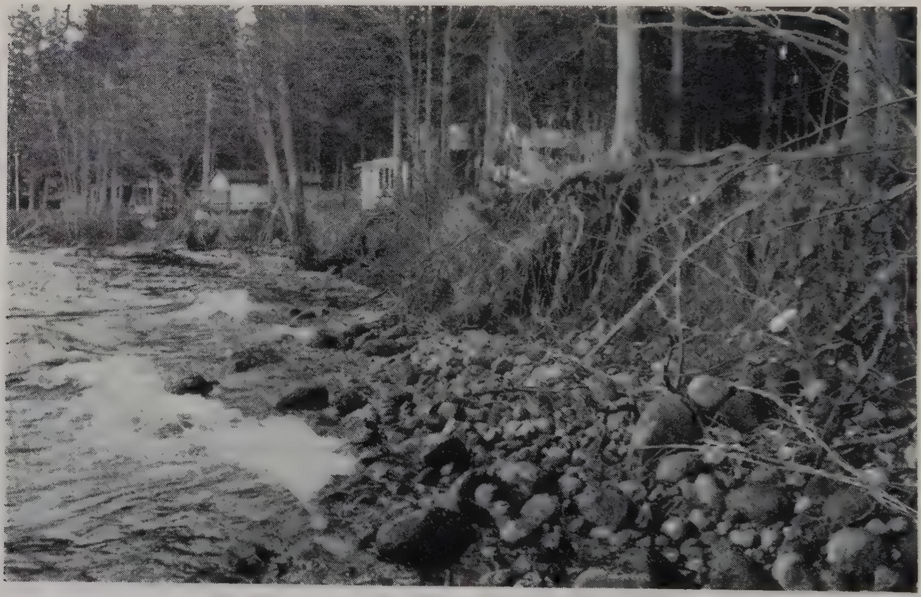


Fig. 104. Low moraine cliff on the northern shore of the Nässja headland. The cottages are situated on the nearshore bottom of the stage preceeding the cutting of the present outlet at Motala. (cf. Fig. 7). Nov. 1963.

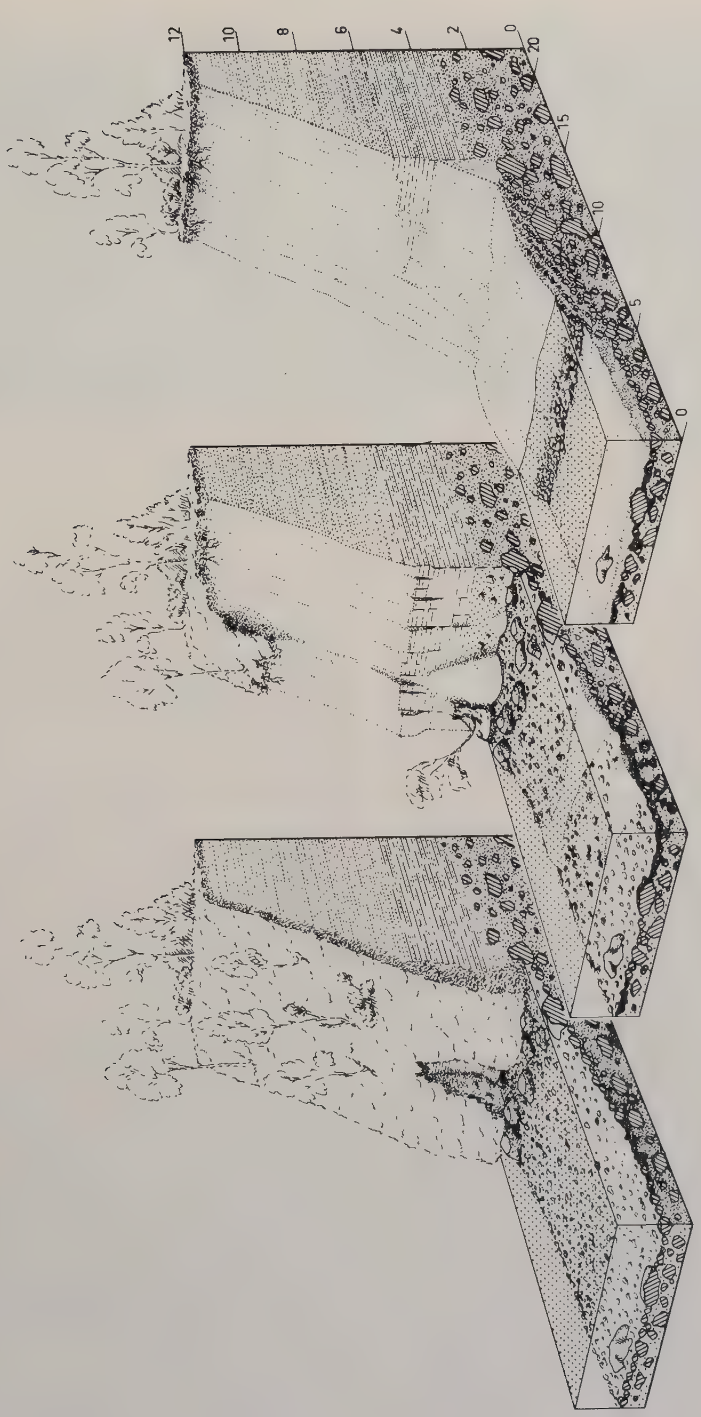


Fig. 105. The morphology of cliffs formed in frictional sediments on cohesive deposits. Based on profiles from the Bankeryd area. Scales in metres.

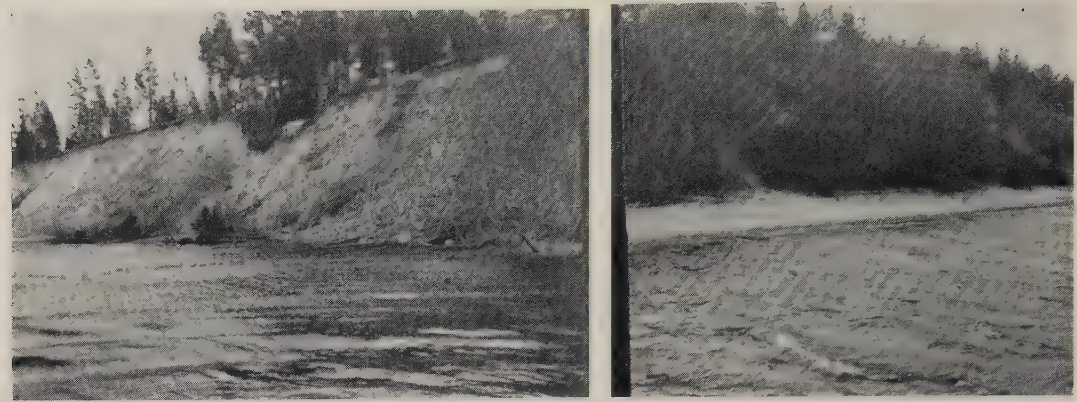


Fig. 106. The morphology on the main downdrift and updrift ends of a concrete landing-stage 300 m south of the river Dumneån in the Bankeryd area. Rock has been dumped south of the stage in order to protect a house on the top of the cliff. Oct. 1962.

Sedimentary frictional deposits underlain by cohesive deposits

This shore type has been illustrated by a triptych of block diagrams (Fig. 105) based on profiles from the Bankeryd area (E in Pl. 1). The soil sequence consist of clayey moraine, silt and sand. The height of the transition from the cohesive moraine and silty sediments to the frictional sand varies from beneath the water level to about 6 m above it (at the Rosenlund cliffs up to 15 m above the water level), but this zone generally lies above the present water level. Thus the direct wave abrasion acts mainly on cohesive material.

The triptych shows the shore in three different stages of erosional activity. The left-hand diagram shows the most common stage. The slope has a vegetation cover, there is some undercutting at the shore scarp, and the beach is narrow or entirely lacking. There may be some residual boulders and blocks pressed up along the scarp by ice pressure. The upper part of the slope has an inclination corresponding to the angle of repose of the sand (26°–34°). The lower part is generally steeper with a somewhat convex profile. This point or beak (Swedish *pynt*), is best developed where there is undercutting, and the hollow is supported by the roots of trees, especially alders, growing at the scarp. This is a common feature, found along many lakes and river banks (Norrman 1955, 1963 *b*). The form is caused by creep in the slope.

In areas of submergence the continuance of this profile depends on a permanent longshore drift. Most shores of this type owe their temporary stability to a delicate balance of longshore drift, and any encroachment upon this drift will immediately cause an erosion. One mistake, sometimes made by man in such areas, is the building of a single solid stage or jetty which will stop the drift and start an erosion on the principal downdrift side (Fig. 106).

Where there is active erosion, the cohesive part stands as a vertical or even overhanging wall, and the frictional part retreats under its angle of repose (the middle diagram of Fig. 105). When wet, even the sandy material can form a steep slope, generally descending in steps. Sand running down corrades the surface of the frictional as well as the cohesive material, forming narrow, vertical furrows (cf. Arnborg 1959, p. 161). The sand is accumulated in talus cones at the foot of the cliff, but is rapidly eroded by the swash.

A continuous retreat necessitates an abrasion of the nearshore bottom. There is always frictional material for corrasion, but the abrasion may cease through too much material being distributed to this zone or through enrichment of the residual material from the moraine forming a stone pavement.

The influence of submergence on the slope of the abrasion surfaces has been discussed in the chapter on wave action in profile. In the shore type described here the outer parts of the nearshore zone and the offshore zone are covered with sand and silt, partly left during the shore retreat and partly transported to these areas from the swash zone. This sedimentation counteracts the effect of submergence on the steepness of the subaquatic slope, and in some areas very flat bottoms are found below ordinary breaking depths (cf. Fig. 71). Owing to the indefinite limit of the abrasion surface, the nearshore bottom is better described as an abrasion plane than as an abrasion terrace.

The right-hand diagram of Fig. 105 shows a shore where a beach has been built up. The retreat has ceased, at least temporarily, and the steep lower part of the cliff has been covered with sand from the upper part. No vegetation is yet found. The first invaders in the lower, wetter, and more fine-grained part will be coltsfoot, alder, and sallow.

The supply of beach material by shore recession naturally depends on the height of the cliff, and also on the percentage of the material which has a grain size large enough to be a suitable beach material. The highest cliffs are found at Rosenlund (30–40 m) and at Sveddudden (25–30 m), and the lowest have a height of 10–15 m. Taking into account the variable part of the cohesive material, it can be estimated that, as a rule, a retreat of 0.5 m gives at least 4–5 m^3 of beach material per metre of shore. A retreat of 0.5 m is the least change which can be expected, when an undercut, steep slope falls down. With regard to the estimated amount of material, it seems that a moderate recession should be sufficient for the development of a lake beach. However, the distribution of the material depends on longshore and outward drift. In shore profiles like the one illustrated in the left-hand diagram, the nearshore zone has been deepened by abrasion, and this zone needs nourishment for the establishment of a beach. Even without direct loss of material the nearshore zone is deepened in relation to the water level by the submergence. This is probably the most serious problem as regards the protection of the submerging sea shores around the world (cf. Bruun 1962). It can be estimated that in southern Lake Vättern about 10–20 m^3 per metre of shore per 100 years are needed to keep the bottom profile down to a depth of 4 m in a constant relation to the water level.

The sediments of the nearshore zone have no large typical morphological forms. Low, subaquatic bars are found along some shallow, somewhat sheltered inundations of the shoreline. The absence of well-developed subaquatic bars can be ascribed to lack of material, and the generally oblique waves approaching the shores from alternate and opposite directions.

The beach has the general characteristics shown in the diagram. It consists of an almost level upper surface, gradually built out during accretion, a beach face with a slope depending on the coarseness of the material, and a step covered with gravel and small cobbles. The narrower the beach is, the coarser is, as a rule, the average grain size, because by drift the finest material is always eroded first.

In the diagram the erosion has been presumed to have ceased because of the development of a sheltering beach. It may also have ceased because of an accumulation of coarse residual material along the shore scarp. In such a case a continuous abrasion in the nearshore zone is still possible. If a repeated recession does not take place until the residual material has been submerged, this recession will be rapid because of the depth in front of the cliff. This should be remembered in areas where there are old shore-protection constructions, built in the form of a boulder and block apron along the shore scarp.

In the central part of Rosenlunds bankar, where the sequence with respect to cohesive and frictional sediments is in principle the same as in the Bankeryd area, the cliff morphology is marked by ground-water drainage, which follows the old erosion channels of the clay surface (cf. p. 36). At the mouth of each channel a cirque has developed with its base in the channel bottom (cf. Figs. 107, 121, 122). The cliff wall is undercut at the outflows of the ground water in the bottom of the channels (Fig. 108), and in this way the cirques have been formed. The material from the walls of the cirque is brought as a mud-flow to the beach during periods of large precipitation and during the thawing period in spring. The flow cuts a furrow from the cirque, and on the beach the material spreads in an alluvial fan. When saturated, these fans are easily abraded by the swash.

In the western part of the cliffs, where there is no continuous clay surface, there are scars in the cliff wall of another kind. Here

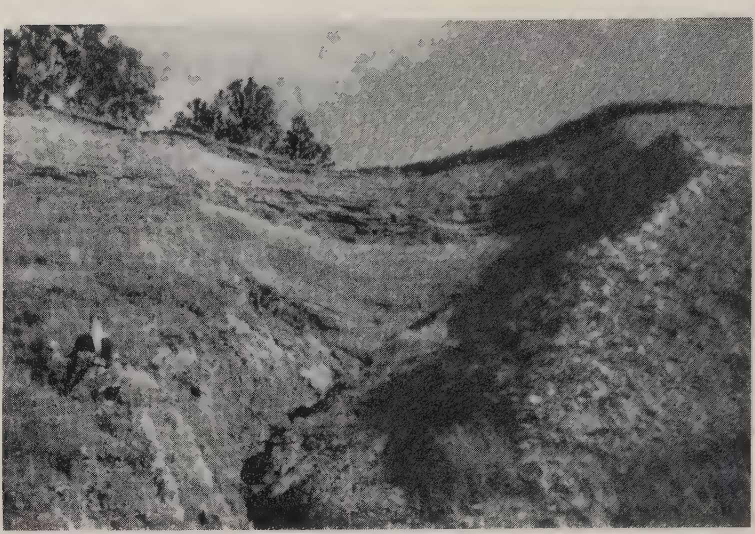


Fig. 107. Cirque in the Rosenlund cliffs (*A* in Fig. 74). The rill in the foreground is cut in cohesive sediments. Aug. 1962.

the indentations are deeper, and at the top of each one a ditch ends (cf. Fig. 120). Already in 1795 the cliff line had a configuration in conformity with the present one. The former pasture land had probably a short time before been turned into tilled fields (cf. Allvin 1839). Thus these scars are ravines formed by the concentrated superficial drainage system of the ditches. It is indicative that in the central part there is only one ravine (cf. Fig. 122), and at this locality there was a small farm in 1710 with tilled fields at the edge of the cliff. There is still superficial drainage through this ravine. East of it the surface of the plateau dips from the cliff, and there no ravines have developed.

Besides the Rosenlund cliffs and the Bankeryd area cliffs formed in frictional sediments

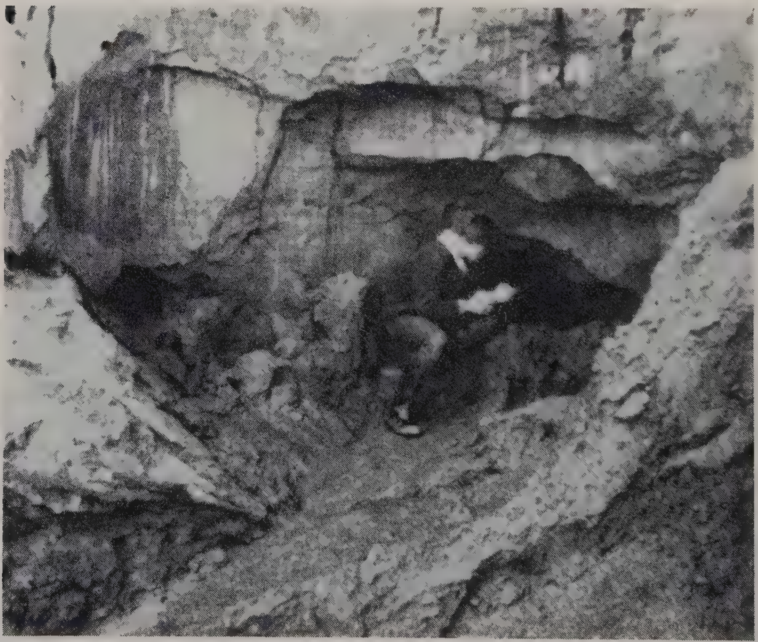


Fig. 108. The bottom of a cirque with ground-water flow. Note the sharp discontinuity (2 m to the left of the man) between the sloping clay surface of the old erosion channel and the fill of fine glacifluvial sand. Aug. 1962.

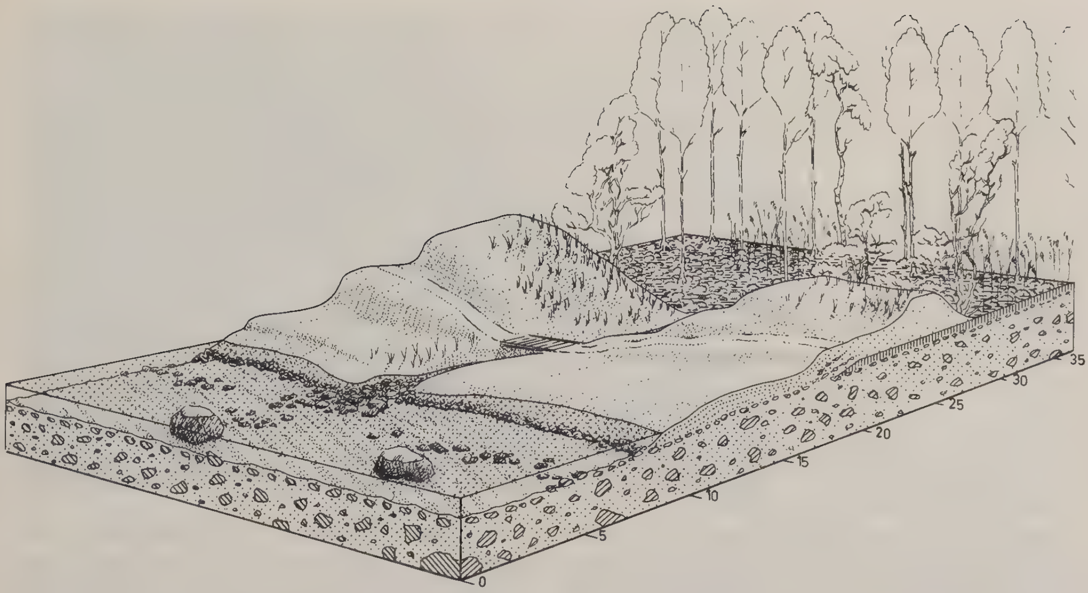


Fig. 109. Morphology of a dune-ridge shore west of the Flisen shoal. The swamp is drained by a small rill.

on cohesive deposits are found north of the Vilhelmsro area (D in Pl. 1) and north of Bankeryd up to the latitude of the northern end of Visingsö island (Skämningsfors). In the former area most of the cliff is covered with grass and bushes, but in some scars varved sediments are exposed from the shoreline to about 4 m above the water level, and above that there is 2–4 m of a non-stratified sandy–silty deposit. In the area north of Bankeryd there are also only isolated scars, and the main part of this shore is in the stage illustrated by the left-hand diagram of Fig. 105. The cohesive sediment is mainly covered by stratified sand.

Flat shores

Shores with a dune ridge

On flat, morainic shores the wave energy is dissipated over a large area. Sand and fine material are washed out from the uppermost layer, but a thin layer of residual material is enough to stop further abrasion. No cliff is formed, and the rate of shore recession is about equal to or slower than the rate of submergence. The zone between high and low water is broad, and sand washed out from the moraine will be exposed to the winds, when the water is low.

The block diagram of Fig. 109 is based on profiles from the shore west of the Flisen shoal (J in Pl. 1). In this area the land nearest the shore is very flat and low, and the submergence during the last 500 years has been about 0.1 m. The flat, nearshore bottom consists of pavements of residual material, patches of naked moraine, and low separate bars of sand. Scattered boulders and blocks project above the water.

Along the shore there is a sand beach and a dune ridge. In the geological map this ridge is classified as a beach ridge formed by material thrown up from the lake (Blomberg 1906, p. 24). In this area there is no predominant direction of longshore drift. In the long, shallow inundations of the shoreline there are accumulations at both ends. The eolic material accumulated has been blown from the beach and from the nearshore bottom at periods of low water. Behind the dune ridge is a swampy zone with alder trees and birches, and in wetter parts a vegetation of reeds (mainly *Phragmites communis*) and sedge. The characteristic feature of this shore is the single and coherent dune ridge. It has gradually covered the organic deposits of the swamp, which are found both under the front and the base of the dune (Fig. 110). Under the organic deposits there is an erosion pavement, which was formed during

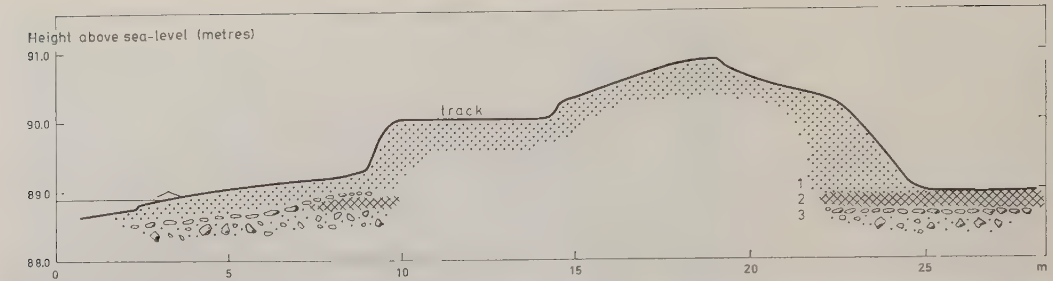


Fig. 110. Profile of dune ridge on the shore west of the Flisen shoal. 1, dune sand, 2, organic swamp deposits. 3, wave-washed moraine. The moraine was abraded during the stage of emergence before 7 700 B.C., and during the present stage of submergence the dune has advanced in over the swamp, which it dams.

the stage of emergence before 7700 B.C., and then comes the moraine.

Diving operations carried out in the offshore zone in a search for old telmatic deposits and remnants of submerged vegetation yielded no results. The bottom is covered with a stone pavement and patches of coarse sand. Any organic material has been washed away during the slow submergence.

The area described is the most southerly one with a dune formed directly on the beach. Further to the south forms of blown sand have only been accumulated on the top of sandy cliffs. The best-developed dune ridges of this type are situated on the Svedudden promontory (cf. Fig. 10) and on the Rosenlund cliffs (cf. Fig. 123).

In the northern section and in the Motala bay beach dunes are found west of the small northern anvil-shaped head of the Nässja peninsula, at the eastern end of the beach (cf. Fig. 74), on the south-eastern end of the land north of the Motala bay (Lilla Hals, north of N. in Pl. 1), on the southern shore of Stora Röknen island (P in Pl. 1), and on Rissnäset north of Karlsborg. The dunes in these areas are of a hummocky type, and in the first three places high (up to 6 m).

The dunes of the Nässja beach have the character of blow-out dunes (Schou 1945, Fig. 65). They dam a small swamp with birch trees. The vegetation of the dunes is spruce. At Lilla Hals the central part of the dunes has been quarried. In the walls of the pit there are several horizons of charcoal. Old Scotch pines grow on the remaining dunes. From the lake the dunes of Stora Röknen island appear as a high wall crowned with pines, but, landward of this front ridge, the morphology is more irregular. The dunes on Rissnäset are also of the hummocky type.

In the north-eastern part of the Motala bay there are naturally forested hummocky dunes inland of a shore scarp with a height of 1.5–2 m. These dunes and the other large dunes in northern Vättern are undoubtedly very old. The dunes of the last-mentioned area and the dunes of Rissnäset are situated in wide areas of glacifluvial sand. They probably started to be formed when the present outlet at Motala was cut and the lake level was rapidly lowered. By this lowering large areas of the former nearshore bottom were exposed to the winds.

In many areas sand has been blown from the beaches without forming any real dunes; a rather thick top layer of eolic sand is found, for example, on the beach ridge of the outer end of Enebågen (cf. Fig. 29) and in the bay north-east of Vadstena (cf. Figs. 111 and 112).

The concentration of the beach dunes to the northern part of the basin—the area of emergence and slight submergence—is in part dependent on the distribution of glacifluvial sand, but it also undoubtedly reflects the conditions of emergence and slow retreat of the shores. The close relationship between the formation of beach dunes, land uplift, beach ridges, and glacifluvial sand in the archipelago off Luleå in the Gulf of Bothnia has been studied in detail by Enequist (1944).

The coherence of the dune ridge west of the Flisen shoal and its juvenile appearance can be ascribed to the thick vegetation behind the ridge, which prevents the formation of blow-out dunes, and to the slow advance of the dune at the same rate as the shore retreat.

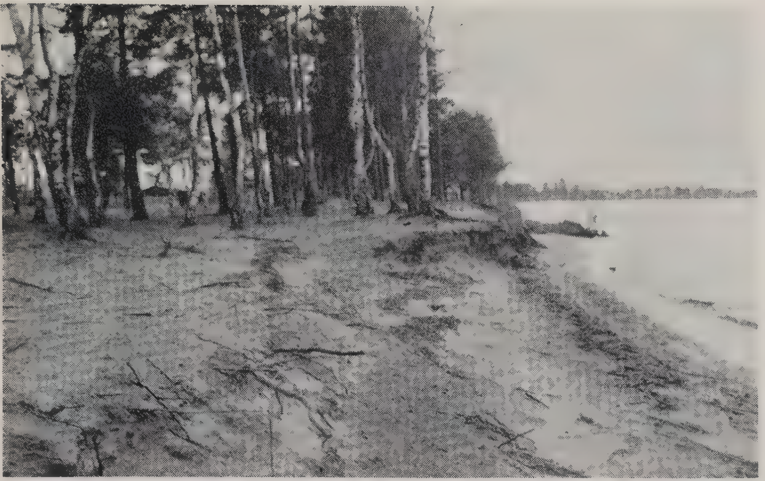


Fig. 111. Abrasion scarp in the bay 3 km NE. of Vadstena. Note the drifted sand and the flat surface. Nov. 1963.

Sandy shores with an abrasion scarp

Low, sandy areas of the northern section and the Motala bay consist of glacifluvial deposits or beach ridges formed at a higher water level. The former are covered with fir trees, the latter generally with grass and bushes.

Where there has been a shore recession in these areas, a shore scarp has been formed. The top layer is held together by the roots of the vegetation, and on this account there may be some undercutting before the upper parts fall down. Because the sand is immediately available, a beach is formed as soon as the momentary erosion ceases. As the source is already sorted, it is difficult to distinguish the abrasion surface from the deposited beach material.

The morphology of the nearshore zone is different in areas of erosion and deposition. As was pointed out in the chapter on wave action in profile in the Vättern basin, general abrasion of the nearshore bottom takes place in areas where there is a pronounced longshore drift. Where the bottom consists of material transported as bed load, partial erosion by transportation off the shore can always be replaced by a reverse transportation. A system of well-developed subaquatic bars is typical of depositional nearshore zones, whereas in abrasional zones there is a more even bottom.

In the bay 3 km north-east of Vadstena (cf. Pl. 1) there is an abrasion scarp cut in sand (Fig. 111). The shore retreats, but the nearshore zone of the bay can be regarded as a

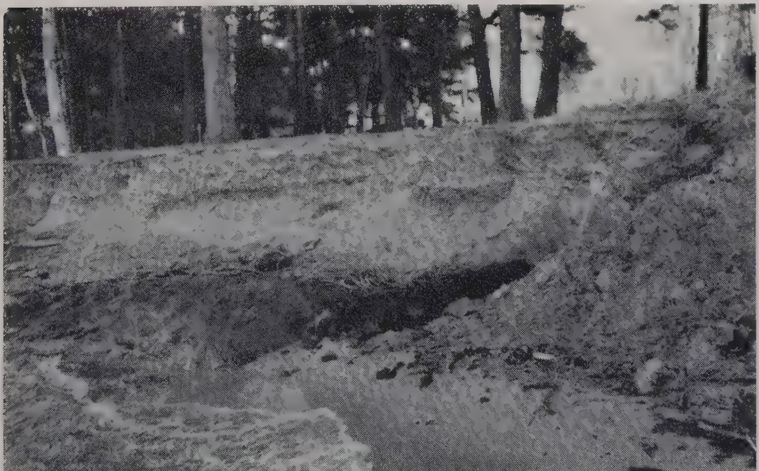


Fig. 112. Abrasion scarp 30 m N. of Fig. 111. Organic lagoon deposit at the base of the scarp. Water stage 88.65 m. Nov. 1963.

depositional one. The shore is used for public bathing, and there have been some minor and unsuccessful attempts to shelter the central parts of the shore with a low wall of concrete blocks and rock fills along the scarp.

The sand of this area comes from marginal glacifluvial deposits, which were accumulated in front of the ice at the standstill in the ice recession. Sand from these deposits was washed down during the emergence stage.

The land is flat nearest to the shore and not higher than 1 to 1.5 m above the present mean water level, but 50 to 80 m from the shore there is a ridge (partly destroyed by a road) with the crest 2.3 m above m.w.l. Behind the ridge there is a depression. The height of the ridge corresponds to the lake level before the cutting of the outlet at Motala, and this ridge of sand was interpreted by Johansson (1955) as having been formed during the stage of submergence before the cutting of the Motala outlet. The interpretation was supported by a find of organic swamp deposits in the present shore scarp (Fig. 112). However, a preliminary palynologic examination by L.-K. Königsson in 1963 proved the deposit to be of a far later origin because of a high percentage of *Picea*, which in any case did not grow in this area before 1000 B.C. The analysis also shows a strong influence of synanthropic plants (i.e. *Plantago lanceolata, P. major, Hordeum, Chenpodeaceae,* and *Juniperus*) which stresses the late origin. An examination of the upper sand strata also shows that these are eolic, and thus the sand surface cannot be correlated to a lake level as high as 2 m above the present level.

The present beach is low and narrow and, because of the sorted sand, the step is insignificant, but there is a runnel at the swash break-point.

The nearshore zone has a width of more than 200 m. It is extremely shallow; the mean depth is less than 1 m, and for 150 m from the shore the bottom is nearly level. This platform could be looked upon as an abrasion platform, because the position of the shoreline is determined by recession. However, its surface and morphology are not abrasional but depositional. The material of the central part of the bay has not only been added by shore retreat but also by longshore drift. The outer part of the nearshore zone is characterized by the subaquatic bars, which have been discussed in the chapter on the shore profiles (cf. Fig. 55).

In principle, the same morphology with a shore scarp and a flat shallow nearshore bottom is found in the Varamo bay (north-west of Motala), but in this area the beach is far broader (6–8 m). The scarp is cut in sand, which in part may be a primary, glacifluvial deposit, but most of it has been re-worked, probably during the first lake stage of the basin.

Morainic meadow and marsh shores

These shores include all types of low shores characterized by limnic, telmatic and terrestrial vegetation. These shores have not been thoroughly examined, because the author is not a botanist, and they are of minor importance along the open parts of Lake Vättern. The vegetation has been described by Stålberg (1939).

The shore type is exclusively found in sheltered parts of the northern section and the Motala bay. From a geomorphological point of view they are characterized by their generally muddy, nearshore bottom and their lack of a morphology developed by ordinary shore processes, and thus they can be used as indicators of sheltered conditions. Their distribution in Lake Vättern depends on the difference in shoreline configuration between the sections of the lake. The probability of their formation is also greater in areas of emergence, because fine, inorganic sediments and organic material, once deposited off the shore, will be raised to a nearshore position in such areas.

Barrier beaches

The formation of barriers has been discussed in the chapters on wave action in plan and the shoreline configuration of Lake Vättern. In this lake they are exclusively found in the area of submergence, where they have straightened the shoreline at the mouths of submerged outlets and low-lying indentations.

In the Huskvarna bay a series of barriers have been formed at the outlet of the river Huskvarnaån (cf. Fig. 147). According to Weiler (1926, 1936), the whole series has been built out from the western side of the outlet in the area north of the old fortress of Rumlaborg. He says that the first tongue was built out

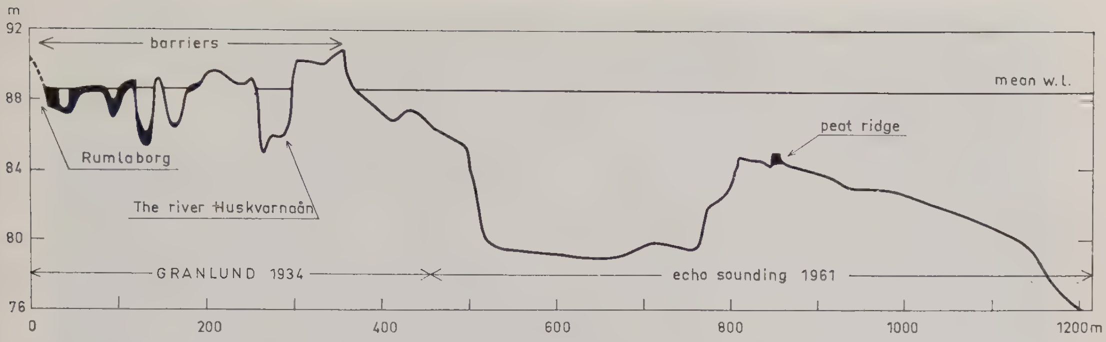


Fig. 113. Profile through the Huskvarna bay from Rumlaborg (cf. Fig. 147) and immediately west of the present outlet of the river Huskvarnaån. Organic deposits black.

from the west, until there was a breakthrough by the river at its base, then another tongue was built out in front of the old one, etc. Thus a whole system of barriers, separated by abandoned channels, was built out. Granlund (1934 *a*) has shown that the height of the barriers can be correlated to the rate of submergence. His profile through the series is reproduced in Fig. 113. The crest of the oldest barrier lies somewhat below the present water level, and this barrier is estimated to have been formed about 700 years ago.

Weiler's interpretation of the development of the barriers postulates that the conditions during the formation of the inner barriers were the same as the conditions during the formation of the present outer barrier, but this is not true. The present outer barrier is built out in line with the shore of the western and central part of the bay. In maps of 1646 and 1647 (Lantmäteristyrelsens arkiv E2: 99, E2: 122) this barrier seems not to have been formed. The river flows straight into the lake, and the shore west of the outlet runs along the pool south of the present railway (cf. Fig. 147) and then towards the northwest. In a map of 1794 (Lantmäteristyrelsens arkiv E.52 1^{4})) there is an outer barrier almost identical with the present barrier (cf. Weiler 1936, Fig. 3, and Granlund 1934 *a*, Fig. 4). However, it can be stated with certainty that the shoreline west of the outlet in the 17th-century map is erroneous.

The pool west of the river is a part of a submerged ravine (now partly filled by dumping), and in this pool the depth is 5 m (Granlund, Fig. 2). If there had been beach-drifting in this area—as presumed by Weiler—the pool would have been sanded up. There is also another fact which argues the erroneousness of the old map; no land west of the river inside the 90 m contour south and *north* of the pool is of recent origin (cf. Fig. 147). This means that the nearest barrier inside the present channel (this barrier lies in line with the submerged ravine) and the still older barriers can hardly have been built out from the western side.

An examination of the profile in Fig. 113 shows that the inorganic sandy bottom of the old channels between the barriers lies above or at most 0.7 m below the synchronous water level, and that their areas are small when compared with the present river channel. As there is almost no bed-load transport in the river, a sanding up of the abandoned channels, followed by a sudden change to organic deposition, seems improbable. It has to be assumed that these channels have never formed part of the main river course.

The combination of unlikely beach-drifting from the west and the positions and areas of the channels makes it probable that all the inner barriers were formed by beach-drifting from the north-eastern shore. The nearshore bottom was gradually built up by bars, and during low water stages swash and beach-drifting formed a new barrier, behind which a channel could be left. The time of the formation of the innermost bar, estimated by Granlund as 700 years ago, agrees well with the carbon-14 dating of the uppermost part of the peat (0.95 m below the present mean water level) of the area G in Fig. 147. It is dated to 590 ± 65 B.P. (Stockholm Carbon-14 Laboratory analysis no. St-1057). This string of peat can be

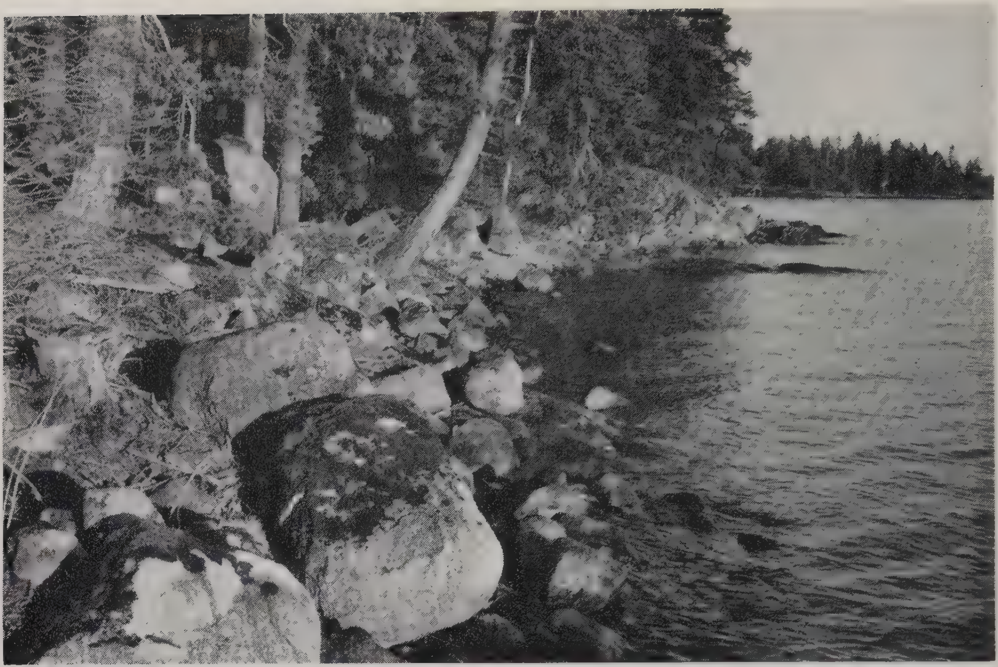


Fig. 114. The shore of the Olshammar esker. Wave-washed coarse glacifluvial material. Ice-pushed boulders in the foreground, outcrops of bedrock in the background. Nov. 1963.

regarded as having been formed in a swamp behind a beach ridge (the peat has been primarily examined by L.-K. Königsson), and it was not until this ridge had been flooded that a drift towards Rumlaborg could start. (The area around the railway station and the harbour has been artificially filled in.)

The barrier of the eastern part of the Jönköping bay is now entirely built over, and it is difficult to separate different ridges from its surface configuration. However, from trenches for pipes, which I examined in 1962 and 1963 (see also Granlund 1934), it is certain that this barrier is not a single unit but a complex formation of different barriers and submerged swamps.

Single barriers are found at the mouths of submerged river valleys and ravines along the western shore. The most northern barrier of this type is at the Svedudden promontory (cf. Fig. 10) and there is another rather large one in the Bankeryd area (cf. Fig. 68). These barriers along the western shore are not broad and complex, because they have been forced to retreat at the same rate as the surrounding shores.

The most northerly of all barriers is situated in the bay of the outer end of Enebågen (cf. Fig. 74). This is formed in a low indentation, and it embays a large swamp. Two other embayments have very narrow barriers, which are better described as beach ridges. One is found in the western end of the southern shore of Enebågen, and the other at the western end of the Erstad bay on Visingsö island (cf. Fig. 76). Thus they are situated in the same position, and have been built up in conformity with the curvature of the surrounding shores.

The material of all the barriers is coarse sand and gravel. The crest of the wave-built part is nowhere higher than 1.2 m above the mean water level, and where the crest is higher, the upper part consists of eolic deposits.

Raised-slope shores

This shore type is developed in frictional deposits with a high content of material coarser than that, which can be effectively transported by the waves. Of the glacifluvial deposits only the eskers fulfil these requirements. Where there is a high percentage of coarse material near the surface of the slope, the waves have mainly a washing effect, and it is difficult to

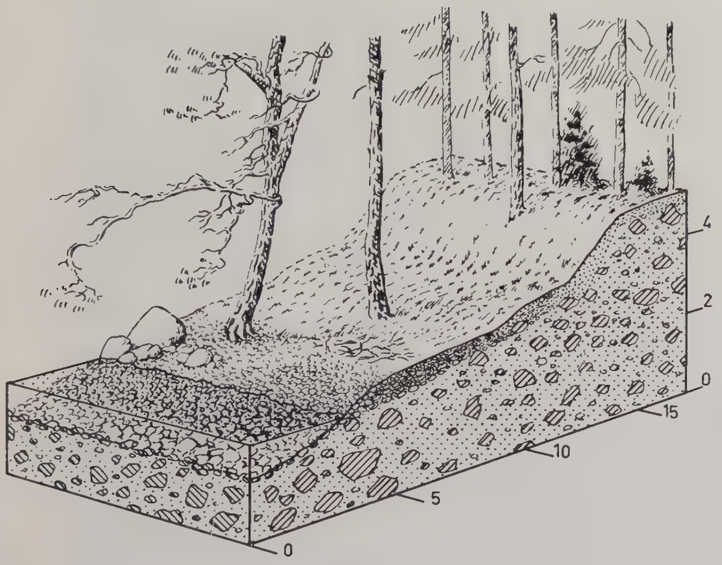


Fig. 115. The morphology of a raised beach slope north of Tokanäset. The upper level at the end of the diagram formed a part of the nearshore bottom before the cutting of the outlet at Motala.

find any distinct morphological forms on the slope. Because of the decelerating rate of emergence the washing is gradually more thoroughly worked out towards the shoreline. The shore is bordered by boulders and blocks (Fig. 114).

The morphology of the nearshore zone depends on the slope and the exposure. If the slope is steep, the nearshore zone is covered with cobbles and boulders. On more gentle slopes the fine material has been washed to about the breaker depth. Nearest the shore there is an irregular stone pavement without a definite lower limit, there are patches of gravel, and in depositional areas the bottom is covered with sand, above which scattered boulders and blocks project.

Where the percentage of the coarsest material is not high enough to rapidly form a covering residuum, the morphology is different. In this case there is often a terrace on the slope, which marks the shoreline of Lake Vättern before the cutting of the outlet at Motala. The present shore is marked by a scarp. In the block diagram of Fig. 115 the scarp is cut in the nearshore part of the higher terrace. This diagram is based on profiles of the shore north of Tokanäset (cf. Fig. 132) on the eastern mainland of the northern section.

The beach is narrow, and its material mainly consists of gravel and cobbles, but there are also scattered boulders and blocks. A large proportion of the cobbles consists of sandstone, which can be regarded as belonging to the lowest stage of the Visingsö Series. No solid rock is exposed in this area or north of it, but it probably makes up the substratum of the shallow offshore bottom.

The nearshore bottom is covered with an erosion pavement down to a depth of about 1.5 m. Further out it is covered with sand, and the bottom is even.

The wave-washed, raised slopes, which are found in the area of emergence, that is, the northern section, do not form a homogeneous type. The block diagram illustrates a profile, which is more marked by wave activity than many of the stretches. Where the moraine is rich in boulders and blocks, as on the headlands of sandstone north of the Motala bay, this activity has only resulted in an enrichment of the coarse material similar to that illustrated in Fig. 114. The generally sharp shoreline with few boulders projecting above the water surface off the shore can be ascribed to the effect of ice pressure. The slopes are generally too steep for the development of ice-pressed ridges. Well-developed ridges have only been found on Tokanäset and on the western shore of Stora Röknen island. In both places the land is very low.

An abraded moss

A moss surrounds the bay 7 km south of Karlsborg, and the shore is cut in the peat of the moss. This shore is a very special one and cannot be included among the main shore types on the lake. However, it is an eloquent witness to the submergence, and the morphology, as such, is also of interest. The shore and the bight form a part of an Army shooting range, and only restricted examinations were possible.

The western part of the moss forms a plateau at 4 to 6 m above the lake. The eastern part slopes towards the bight, and along the shore its surface lies 1 to 2 m above the lake. In the central parts the peat has a thickness of 5 to 8 m, and along the shore it is probably 3 to 4 m (cf. Willén 1926). From the morphology it can be judged that the peat of the sloping part is moving towards the shore. There are coherent cracks parallel to the contours of the slope. These cracks have a width of about 0.5 to 1 m, and they lie 10 to 20 m apart (Fig. 116). Their depth is about 0.5 to 1.5 m. Near the shore there are also cracks in the direction of the slope, and the cracks form a rectangular pattern. This pattern resembles the permafrost polygons of the Arctic tundra. Here the frost is certainly not the primary agency, but it seems probable that the winter frost actively affects the cracks.

The bottom of the cracks has been observed to freeze before the peat at the same level between the cracks. Water continues to ooze out of the walls, and a kind of ice wedge is formed in the cracks. No measurements have been made, but it can be expected that the frost penetrates far deeper into the cracks than into the peat beside them.

The shore front is undercut by the waves, and is dissected by abrasion along the cracks (Fig. 117). The front wall is vertical or overhanging to a depth of 0.5 to 2 m. The bottom of the bight is flat and covered with detritus. About 50 to 100 m from the shore there is a belt of reed.

A comparison between the economic map of 1877–82 on a scale of 1 : 20,000 (Ekonomiska konceptkartan över Skaraborgs län) and the economic map surveyed in 1955–57 and published on a scale of 1 : 10,000 (Ekonomiska kartan över Sverige, 8 E Hjo 6 e Hammaren) indicates a retreat of 30 to 50 m. The shoreline of the older map may be somewhat uncertain. According to Pehrson (1942, p. 13), the retreat in the period 1924–28 was about 10 m, as estimated by local people.

The original distribution of the moss is uncertain. It seems probable that it once covered the main part of the bay, and that it was formed behind a beach ridge that was abraded during the present stage of submergence.

Recent shore retreat

Introduction

Shore retreat by abrasion in southern and central Lake Vättern has long been a matter of economic importance. Although the submergence of this area is a primary postulate of the retreat, the striking activity of the waves has been most observed, because the retreating shores are cliffs.

Valuable information about the shores in the late 18th century and the 19th century is to be found in topographic descriptions by Allvin (1839, 1859). Allvin (*b.* 1777, *d.* 1866) was the senior land-surveyor in Jönköping, and his knowledge of the shores was undoubtedly a consequence of his professional experience. The landowners disliked being taxed for land that had been abraded, and in allotments nobody wanted to have the land nearest the cliffs, and therefore a strip of land along the shore had to be left out. This land was called *sjövrak* (lake refuse).

Because the submergence of southern Lake Vättern was too slow to be understood as a regular natural phenomena, the flooding of low-lying areas in Jönköping was ascribed to artificial damming of the outlet at Motala, and during the 17th century the inhabitants of Jönköping complained to the King about this damming (Björkman 1919, pp. 328–332). After the town had been burnt down in 1612, a new town was built on the low barrier east of the outlet from Lake Munksjön, and it was this area that was flooded. Allvin gives an informative description of this area in connection with a discussion of the retreat of the Rosenlund cliffs east of the bay: "... this bay seems to be more and more filled up, and the houses which because of deficiency of space were considerably built out into the lake now stand

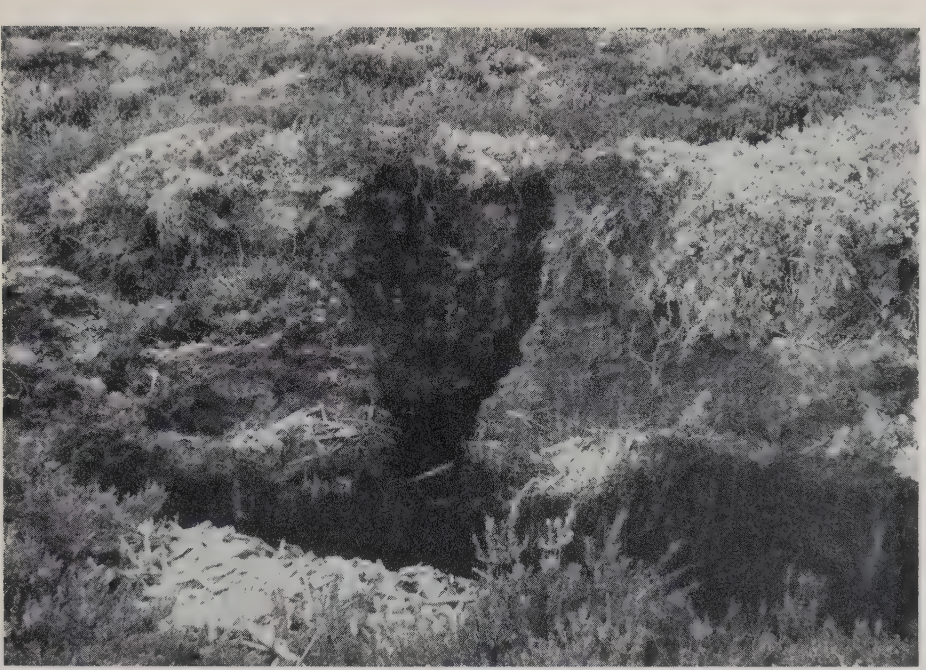


Fig. 116. Junction between 0.5 m wide cracks in a moss on the shore of the bay 9 km south of Karlsborg. Nov. 1963.

on dry land and a beach has also been formed in front of them" (Allvin 1839, p. 37). With the houses built out into the water it is no wonder that they were flooded during ordinary high-water stages, even without artificial damming.

Allvin mentions attempts to estimate the retreat of the Rosenlund cliffs between 1710 and 1798 and between 1799 and 1808 (1859, p. 25), the retreat of parts of the eastern shore of Visingsö island between 1800 and 1835, and the retreat in the Gränna area between 1805 and 1858 (op. cit., pp. 23–24). In the first few years of the present century an investigation on the possibility of shore protection on Visingsö island was carried out (Väg och Vattenbyggnadsstyrelsen) by A. Blomberg, a major in the Swedish Corps of Engineers. His reports were never published, and I happened to find them in private keeping in Gränna. It may be mentioned that the construction of a rockfill apron along the Rosenlund cliffs was discussed as early as 1798 (Allvin 1859).

A map on a small scale (1 : 50,000) of the shore retreat around the southern end of Lake Vättern "since the beginning of the 18th century" has been published by Weiler (1936, Fig. 5). It is based on comparisons with old maps. A similar map of the shore retreat around the northern headland of Visingsö since 1839 has been published by Eckerberg (1951) on a scale of 1 : 30,000. Because of the scale these maps are of little value. However, it should be mentioned that Weiler has discussed the retreat as he found it in the old maps (1926, 1936).

Old land-survey maps on scales of 1: 4000 and 1: 2000 have proved to be reliable for the determination of shoreline changes along rivers (Sundborg 1956, Sundborg and Norrman 1963). However, the jordebok maps (tax records) from the 17th century cannot be used, as they have, as a rule, errors in the shore contours that are too large (Sundborg 1956, p. 247). They have also been found impossible to use in Lake Vättern. It was the arable land that was of fiscal interest, and thus this land was mapped with care. The contours of uncultivated land were often drawn freehand. In this respect we are fortunate in the Vättern area, because there were tilled fields on the edge of the cliffs, and the contour of the maps shows the cliffline. Sometimes the shoreline was also drawn, sometimes not.

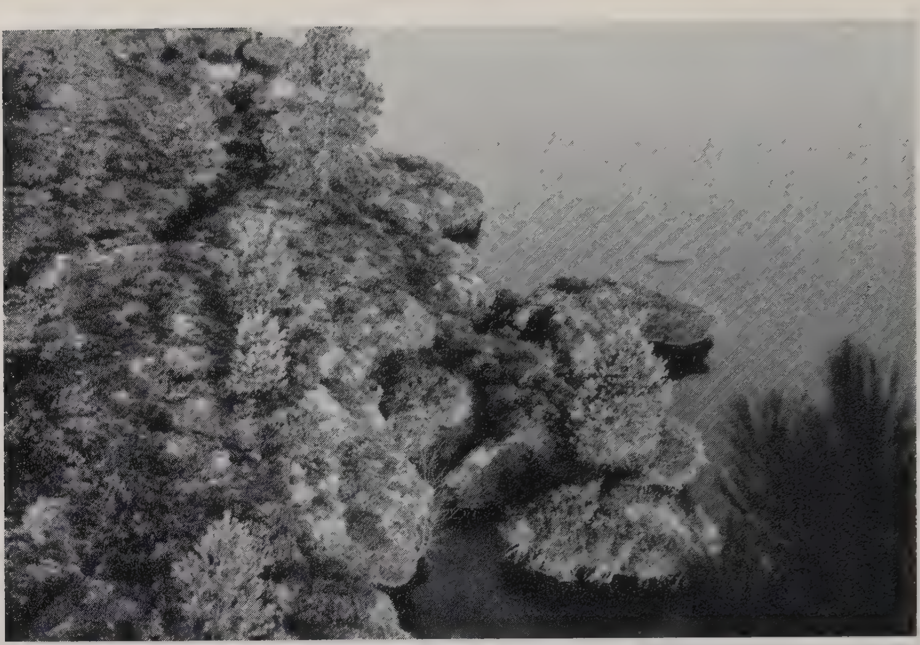


Fig. 117. Shoreline contour of the moss 9 km south of Karlsborg. Nov. 1963.

In 1958 I was asked by the Committee on the Shore Erosion of Southern Lake Vättern (Stranderosionskommittén för södra Vättern, sekr. jordbrukskonsulent Drake, Hushållningssällskapet, Jönköping) to state the rate of erosion and the possibilities of shore protection in the area from Visingsö island to the southern end of the lake, and therefore most of the detailed investigations were carried out in this area. Reports were presented to the Committee in 1961 and 1962. In the area investigated the retreat has been most active, even though there is also appreciable retreat in parts along the shore of Guldkroken and the Ostrogothian Plain. The recent erosional activity along the latter shore has been estimated by Johansson (1955).

Here the retreat of the shores on Visingsö island and around the southern end of the lake will be discussed more in detail, and the rest of the lake will be discussed in its general features. Within the area of interest to the Committee on Shore Erosion all the maps kept in the archives of Lantmäteristyrelsen (National Land Survey Board) in Stockholm, the local land-survey offices in Jönköping, and of the municipal engineer of Jönköping, have been carefully examined.

The first maps, which can be used for comparative studies, are some geometrical surveys from the beginning of the 18th century. However, the contours of these maps are generally not detailed, and this type of maps has to be used with care. The first series of maps, having an accuracy that really permits a detailed comparison with the modern maps, was made in connection with the *Storskifte* (cf. Sundborg 1956, p. 248). *Storskifte* means a re-allotment of land, whereby the narrow strips of the common-field system of each hamlet were gathered into larger fields. The later *Laga skifte* reform, which gave rise to new surveys, was meant to gather all land belonging to one owner into one coherent area. The *Storskifte* of the areas studied took place during 1788–1818, and the *Laga skifte* during 1839–72.

Since the middle of the 19th century several maps have been made in connection with the partition of farms, change of ownership, disputes about boundaries, etc. In the Jönköping–Huskvarna area there have been special surveys for the planning of the towns.

The titles of the maps used and the names of the surveyors have been judged to be of small interest, and as no important differences

Table 18. Summary of the shore retreat (in metres) on Visingsö island according to measurements by Blomberg in 1907

Hamlet	Boundary[1]	Maps (E. 127)	Year	Dist. to landmark	Retreat	Average annual retreat
Ed[2]	5^2-5^3	7^4	1800	126.5		
			1907	118.0	8.5	0.08
Ed	10^1-4^6	7^4	1800	86.5		
			1907	74.3	12.2	0.11
Ed	10^1	7^4	1800	234		
			1907	210.5	23.5	0.22
Ed	10^1 (landing place)	7^6	1839	17.8		
			1907	2.0	15.8	0.23
Torp	7^1 A (landing place)	28^4	1850/51	17.2		
			1907	0	17.2	0,30
Lundby	1^4 (landing) place)	24^3	1852	8.9		
			1907	1	7.9	0.15
Husgärde/Rönäs	1^5-3^1	?	1880	274		
			1907	269	5	0.19

[1] Letters according to the economic map of 1953.
[2] Northern end of the island, all the others from the western shore.

have been found between the original maps and contemporary copies, the old survey maps are, as a rule, only referred to by year and by the letters in the archive register of Lantmäteristyrelsen.

Shore retreat on Visingsö island

In one of the previously mentioned works by Allvin he states that a survey had recorded a retreat of 68 Swed. ft (20.2 m) at the hamlets of Erstad and Ed between 1800 and 1835 (1859, p. 23), hence an annual retreat of 0.59 m. This shore is situated on the western side, "where the island is most narrow".

Some estimations of shore retreat are found in Blomberg's reports. He had no new map to compare with and did no surveying, and thus he could only estimate the retreat along boundaries with landmark stones and at landing places which had been mapped as *sjövrak* in the middle of the 19th century. His results are summarized in Table 18. He found very few landmark stones in relation to the number marked in the old maps, and in my own search for these stones I did not find one single stone on its original site.

The maximum annual retreat, according to Blomberg's measurements, was 0.3 m, the minimum retreat about 0.1 m, and the average value, according to the figures of Table 18, is 0.18 m. It may be mentioned that the landowners reported far higher figures to Blomberg, based on their own estimations.

Old people on Visingsö have told me that Blomberg levelled shore profiles, and these were thought to be most valuable for a determination of the retreat during the last 50 years. It was also found that Blomberg had levelled shore profiles for every 50 m around the island (1905–1907, dated 1911), but none of the profiles is referred to any kind of landmark in the direction of the profile, and thus they are of no value.

The photogrammetric mapping of Visingsö island in 1953 (published 1956) made new comparisons with the old maps possible. In this work I was assisted by Miss Ella Nilsson (1957), who investigated the shore retreat of the southern and south-western shores. The photogrammetric maps on a scale of 1 : 10,000 were enlarged to 1: 4000, and thus the comparisons were made on the original scale of the old maps. From comparisons between the boundaries of the maps I found that 4 m is the least change of the cliff line that can be recorded as a certain retreat. The result of the comparisons cannot be graphically reproduced as a whole on an appropriate scale. Two examples are shown in Figs. 118 and 119, one from the northern end of the island and one from the western shore in the hamlet of Husgärde (north of G in Pl. 1).

The results of the comparisons are summarized in Table 19. It was found that on the entire eastern shore south of the Erstad bay the retreat has been less than 4 m during the

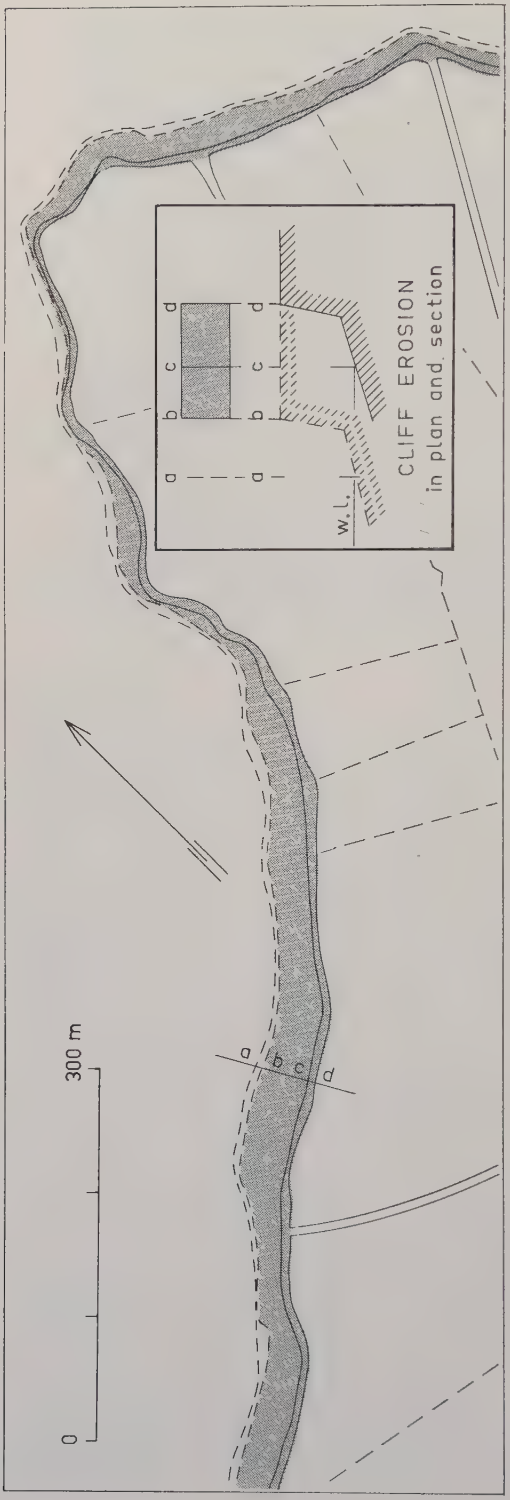

CLIFF EROSION
in plan and section
a
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Fig. 118. Retreat of the cliffline on the northern end of Visingsö island in the period 1880–1953.

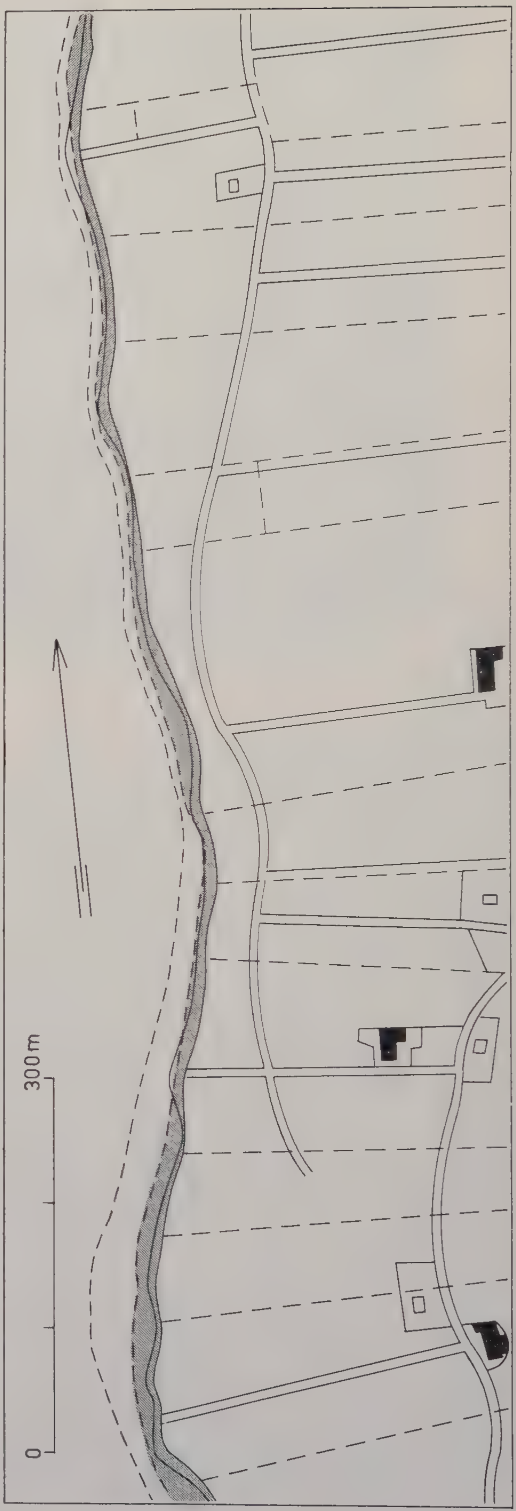


Fig. 119. Retreat of the cliffline in Husgärde on the western shore of Visingsö island in the period 1850–1953.

Table 19. Shore retreat (in metres) on Visingsö island, according to map comparisons.

Locality	Period	Maps (E. 127)	Maximum retreat	Average retreat	Average annual retreat
Ed, from 2^3 to 3^3B	1802-1953	7^5	40	20	0.13
Ed, from 3^3B to 4^5	1800-1953	7^4	45	25	0.16
Ed, from 4^5 to 1^3	1800-1953	7^4	28	15	0.10
Rökinge	1767-1851	24^1	10	8	0.10
	1851-1953	30^4	20	8	0.08
Lundby	1767-1850	24^1 24^3	5	4	0.05
	1850-1953	24^3	10	8	0.08
Husgärde	1767-1850	24^1 15^1	12	4	0.05
	1850-1953	15^1	25	12	0.12
Rönäs	1767-1953	23^1	36	15	0.08
Asby, north of Bengtsgården	1838-1953	3^2	15	8	0.07
Asby, south of Bengtsgården	1838-1953	3^2	8	< 4	< 0.03
Näs, the western shore	1767-1835	23^1 23^2			
	1835-1953	23^2	Practically no shore retreat		
Näs, western part of the southern bay	1767-1835	23^1 23^2	8	< 4	< 0.06
	1835-1953	23^2	8	< 4	< 0.03
Näs, eastern part of the southern bay	1767-1835	23^1 23^2	8	4	0.06
	1835-1953	23^2	20	15	0.13

last 150 years. As can be seen from the table, the average annual retreat has been far less than the figure reported by Allvin and also generally less than Blomberg's figures. However, the figures of maximum retreat within the different stretches show that, at a certain locality, the annual retreat may be very high for some decades. These are the places noted by the local people, and thereby the general rate of retreat may be overestimated.

The rapid changes in the rate of shore retreat are illustrated by the shoreline changes along the Visingsö airfield, situated on the north-eastern shore, west of the Sandudden ness. This area was mapped in 1934 on a scale of 1: 1000 and photographed in 1951. The retreat has been recorded in relation to fixed air-lights situated along the shore and in Table 20 numbered I to VII from west to east. This table clearly demonstrates the large variations in space and time owing to changes in the distribution of beach material, water-stage variations, and variable wind conditions. The retreat is not continuous. When a cliff has collapsed through undercutting, it will take some time before this material is removed. During periods of a low water stage the shore bench, and not the cliff, is abraded.

Of a special interest is the southern part of the south-western shore of the island, which is protected by a shingle beach, and where there has been almost no retreat of the moraine cliff since at least the middle of the 18th century (Table 19: Näs, western shore, Asby south of Bengtsgården). During my investigation of this shore a retreat had just started at its northern end (Asby $4^{4, 5}$), and it was found that the soil of the gentle slope

Table 20. Shore retreat along the Visingsö airfield in 1934-62.

Air-light nos.	Average annual retreat (metres)					Total retreat
	1934-51	1951-61	1957-61	1961-62	1934-62	1934-62
I	0.59	0.02	0.06	0.33	0.38	10.5
II	0.35	0.15	0.33	0.53	0.29	8.0
III	0.35	0.07	0.30	0.13	0.24	6.8
IV	0.70	0.00	—	0.00	0.43	12.0
V	0.24	0.01	—	0.00	0.14	4.0
VI	0.00	0.04	—	0.33	0.03	0.7
VII	0.00	0.15	—	0.00	0.05	1.5

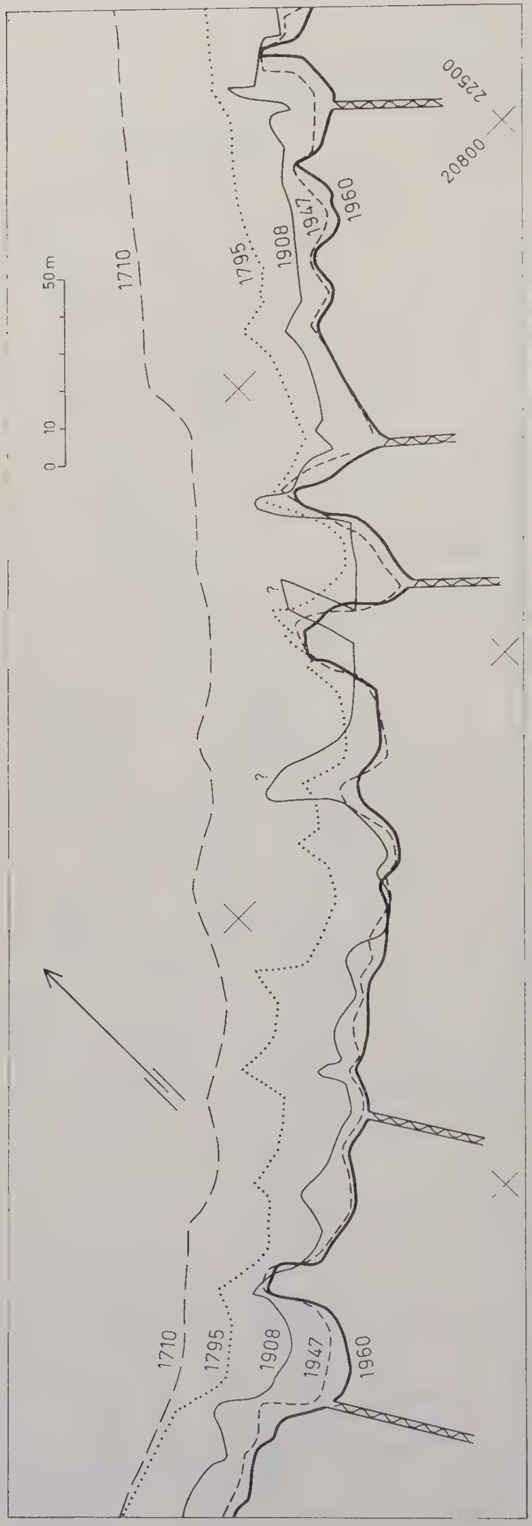


Fig. 120. Retreat of the cliffline in the western part of the Rosenlunds bankar (area IV in Fig. 74) from 1710 to 1960.

rested on a sand beach. Just above the sand there was a layer with iron slag, charcoal and bones. These deposits had obviously been dumped at the foot of a cliff, then buried below talus from the cliff wall, and since then there had been no retreat of the shore. The age of the charcoal was determined at the Uppsala Carbon-14 Laboratory (Olsson 1959) to be 535 ± 70 years BP. Thus it can be estimated that this shore has been stable since about the beginning of the 15th century. During this period the submergence of the area has been about 0.45 m. The dating of this shore is of practical interest, because it shows that the shore retreat can be stopped for a long time by shingle deposits.

Shore retreat at the southern end of the lake

From Weiler's papers (1926, 1936) it is known that map comparisons can be made in this area and that the retreat is most rapid along the Rosenlund cliffs. The upper contour of these cliffs has in modern times been surveyed twice on a scale of 1: 1000 in the coordinate system of Jönköping, first in 1908 and then again in 1947. The same survey was made in the Vilhelmsro area on the western shore north of the Jönköping bay in 1908 and in 1956. Copies of the manuscripts were placed at my disposal by Mr Norsander, the municipal engineer.

With regard to the rapid retreat of the Rosenlund cliffs it was thought worth while to make a new survey in this area, and this was done in 1960–61 by one of the students at the Institute of Geography in Uppsala, P.-A. Björklid (1962), who also studied the relation between the vegetation and the morphology of the cliffs. The retreat in the Rosenlund area will first be discussed.

The Rosenlund cliffs

Weiler judged the oldest large-scale map from 1710 (E. 141 4[1]) not to be accurate enough for a detailed map comparison (1936, p. 26), and he seems not to have noticed a copy of this map from 1786 with the cliffline surveyed in 1795 added (archives of the municipal engineer, Jönköping F VI 40/2). On this map there is a note: "Here runs the shoreline in 1795, surveyed by J. Montelin". Montelin was

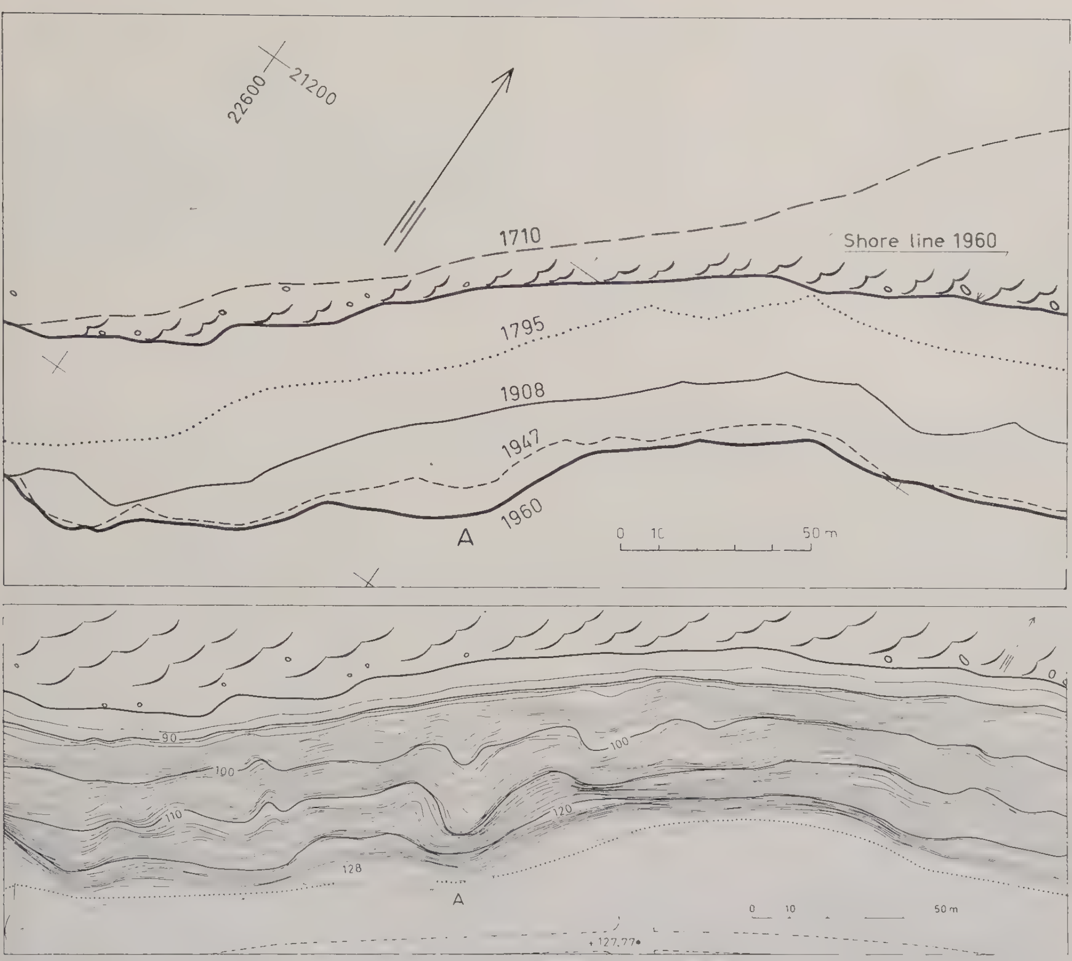


Fig. 121. Morphology of the eastern part of the Rosenlund cliffs (area V ni Fig. 74) with well-developed cirques. The upper figure shows the retreat of the cliffline in the same area.

the surveyor in Jönköping at this time. It is certainly not the shoreline that is drawn, because the line has the characteristic configuration of the upper cliffline formed by ravines and cirques. This line is very detailed, and there is no difficulty in identifying the present ravines (cf. Fig. 120). However, no new landmarks were added in this survey of 1795, and it was first found impossible to fit the landmarks of the 1786 copy or the identical ones of the original map from 1710 to a modern map with landmarks from 1904 (archives of the municipal engineer, Jönköping F VI 270/1). With the help of an optical pantheograph it was found that the map of 1710 was on a somewhat different scale than that noted on the scale line (1 : 4,000), but that the contour of 1795 had been drawn according to this scale. When the contours had been corrected to the same scale, the maps could be connected to the modern ones.

With regard to the retreat and the morphology the cliffs can be divided into a western and an eastern part, separated by a forested section at the step of the shore contour (between area IV and V in Fig. 74). It has not been possible to reproduce the map comparisons along the full length of the cliffs on an appropriate scale, so they have been exemplified by three areas (Figs. 120—122) with the clifflines of 1710, 1795, 1908, 1947 and 1960.

The area shown in Fig. 120 is from the

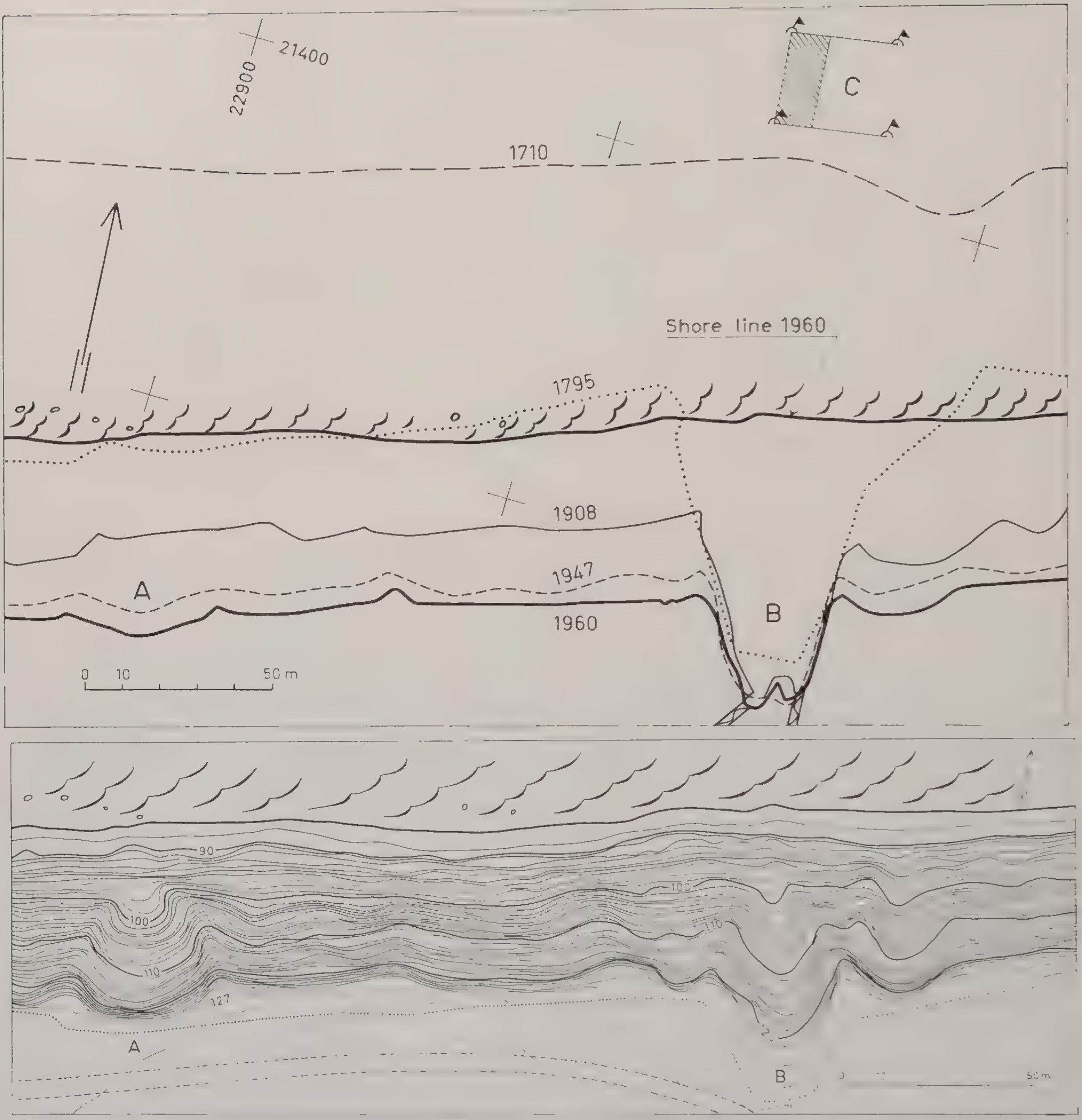


Fig. 122. Morphology of the eastern part of the Rosenlund cliffs (area VI in Fig. 74). There is one large cirque (*A*) and two other cirques at the sides of the ravine (*B*). The upper figure shows the retreat of the cliffline in the same area from 1710 to 1960. (*C*) marks the area of detailed bottom investigations. The mapped area of Pl. 3. is shaded.

western part, where there are ravines connected to open ditches. The cliffline of 1710 is much more even than the others. This may in part be due to a simplification, but it seems most probable that there were no ditches at this time, when the plateau was still grazing land belonging to the Crown, and therefore no deep ravines. It may be noted how well the contour of 1795 fits the configuration of the modern maps, but the ravines are not as deep as to-day.

A photogrammetric map of the eastern part of the cliffs on a scale of 1: 1000 was constructed by Rikets Allmänna Kartverk (Geographic Survey Office of Sweden) in 1962 on aerial photographs from 1960. Fig. 121 shows the area immediately east of the forested section. In the map of the retreat the shoreline of 1960 has been drawn in order to facilitate comparisons with the contour map. In this area there are no ditches and ravines but

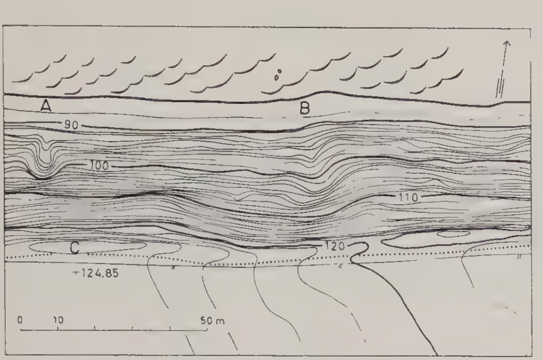


Fig. 123. Morphology of the eastern part of the Rosenlund cliffs (area VII in Fig. 74). There are two embryonic cirques (*A* and *B*) and a dune ridge at the top of the cliff (*C*).

cirques (cf. Fig. 107). Even here, the cliffline of 1795 follows the contour of 1908 well, and so does the 1710 cliff, except in the eastern part, where it deviates outwards.

Fig. 122 shows the cliff further to the east (cf. Fig. 11). In this area the only ravine in the eastern part is found (B), and it may be noted that it is connected to ditches. There are also cirques (A, and at the sides of B). The map comparisons indicate an extremely rapid retreat between 1710 and 1795. As regards the position of the cliff contour of 1710, it should be mentioned that, as connected in Fig. 122, it runs just off the present edge of the eastern end of the cliffs, where, because of coarse boulders, a very small retreat can be expected (cf. Fig. 12). Thus the contour seems to be correct at that end, but it may, of course, be erroneous in the area shown, which lies 600 m further to the west. In this respect the shallow indentation of the 1710 contour and the deep ravine of 1795 are of interest. If there was a deep ravine in 1710, this should certainly have been marked, because there were some tilled fields just on the cliff edge along the eastern part of the indentation and also a farm-house (these are the only tilled fields along the entire cliff in the map of 1710).

The direction between the indentation and the bottom of the present ravine coincides with the estimated direction of the erosion channels in the clay of the lower cliff (cf. p. 36). Thus it seems probable that the indentation marks a cirque, and that ditches were dug towards this from the fields, and then a ravine developed. As can be seen from the map, there has been almost no headward erosion in the ravine since 1795, but a considerable retreat of the surrounding cliffs. The headward erosion was stopped by the dumping of brushwood and stones.

Allvin (1839, p. 37) says that the cliff had retreated 101 m (340 Swed. ft) in 92 years. This figure must refer to the maximum retreat at the ravine of the eastern part of the cliffs. In 1859 (p. 25) he says that, from comparisons between the map of 1710 and a survey of 1798 (possibly the survey noted as having been made by Montelin in 1795), the loss of land along the cliffs was 101,200 m² (20.5 tunnland). This value means an average retreat of about 35 m, which agrees well with my comparisons (0.40 m/year). According to Allvin, there were also surveys in 1799 and 1808, which showed a loss of 28,400 m² (5.75 tunnland). These surveys at the end of the 18th century and the beginning of the 19th century were undoubtedly made because the Rosenlund area had been let to Baron G. M. Posse, who undertook to prevent further retreat by dumping brushwood and stones and by planting suitable vegetation (cf. Bjurulf and Weiler 1953, p. 15).

It seems probable that the contour of the 1710 map does not deviate much from the cliffline of that time. This means that during the first half of the 18th century the shoreline passed the area with the specially investigated bottom morphology marked in Fig. 122, where the depth by abrasion is now between 2.3 and 4 m.

During the period 1795–1908 the retreat was much slower. In the eastern part the average retreat can be estimated to be 24 m (0.21 m/year), and in the western part 11 m (0.10 m/year). According to the maps, there was no retreat in the forested section. This can be judged as correct, because Björklid (1962)

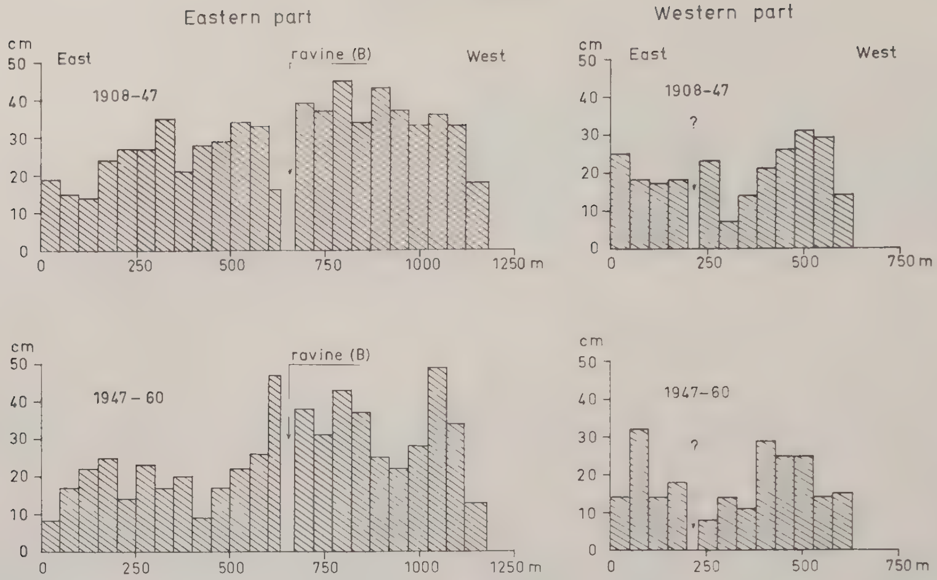


Fig. 124. Distribution of the annual shore retreat along the Rosenlund cliffs in the periods 1908–47 and 1947–60.

identified trees on the cliff that were about 110 years old, which is thus the minimum period of stability.

The retreat during the 20th century can be stated more precisely. The total loss in the period 1908–47 was 17,840 m² (460 m²/year), of which 13,210 m² (340 m²/year) was in the eastern part and 4630 m² (120 m²/year) in the western part. In the period 1947–60 the total loss was 5090 m² (390 m²/year), of which 3710 m² (290 m²/year) was in the eastern part and 1380 m² (110 m²/year) in the western part.

The annual retreat for every 50 m of the cliffline during the two periods is illustrated in Fig. 124. Near the eastern cliff end the retreat is small, because of the high content of boulders and blocks. In the eastern area the retreat was most rapid west of the ravine. The maximum annual retreat of the first period is 0.45 m and in the second period 0.48 m. In the western area the average annual retreat is smaller and also the maximum retreat (0.30 m).

The distribution of annual retreat along the cliffs is more uneven in the second period, a fact which can be ascribed to its shortness. It may be noted that, where the retreat was rather large during the first period, it is comparatively high during the last and vice versa. Although the total annual loss is smaller during the latter period, a general decrease in the rate of retreat cannot be predicted.

The volumes eroded in the period 1908–60 have been estimated on the assumption that the height of the cliffs has constantly been the same as in 1960, that the slopes of the cliff wall and of the nearshore bottom were the same in 1908 as in 1960, and that the base of abrasion lies at a depth of 3 m below the mean water level. This estimation gives a total volume of 826,000 m³ or on an average 15,900 m³/year.

The western shore and the Huskvarna bay

In the small-scale map by Weiler (1936, Fig. 5) a retreat of about 50 m “since the beginning of the 18th century” is marked along the western shore north of the Jönköping bay. I have tried to make use of the old maps of this area, but the connections between these and modern maps were found to be too uncertain for a reliable determination of the retreat. The only retreat on this shore that can be stated with certainty is the one obtained by the surveys of the Vilhelmsro area in 1908 and 1956. The annual retreat calculated at 50 m intervals is shown in Fig. 125. The total loss along the shore surveyed (890 m) was 13,640 m², and the average annual retreat was 0.32 m. The retreat was most rapid in the southern

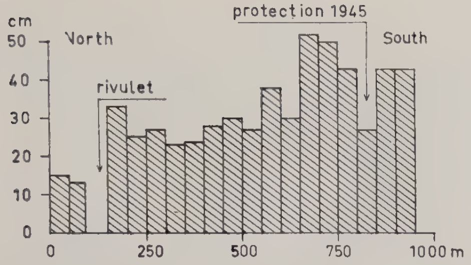


Fig. 125. Distribution of the annual shore retreat in the Vilhelmsro area in the period 1908–56.

part, especially around a sewage-disposal plant situated on the cliff. Here it was about 0.4–0.5 m/year. Just in front of the plant there is a concrete wall along the shoreline.

South of the Vilhemsro area down to the Jönköping bay the shore is now partly protected by rock fills and concrete landing places and walls.

The retreat in the Huskvarna bay has been paid much attention in the topographic literature, because the western part of the bay was the site of Sanna parish church, which was probably abraded at the beginning of the 18th century (when the church had very likely already been demolished). The history of the parish and the church has recently been published by Silfving (1960). In the 1760s there was still a part of the churchyard left on the shore (Silfving, p. 28). In a map from 1646 (E2: 123–124) the church is marked as situated on the shore about 400 m west of the river Huskvarnaån. The shoreline of this map must be regarded as very uncertain. It has been used by Weiler, together with a map from 1799 (E. 141, 16^{1}), for the determination of the retreat until 1930 (1936, Fig. 3). According to his comparison, the retreat in the middle of the western part of the bay was about 45 m (about 0.30 m/year) in the period 1646 to 1799, and about 22 m (about 0.17 m/year) in the period 1799 to 1930. The figures can be judged as reasonable, and from my own comparisons I have found the retreat to be the same during the latter period, but I have not found it possible to use the map from 1646.

The present retreat is very small, mainly because of rock fills along the shore. The eastern accretional part of the bay has been discussed in the chapter on wave action in plan and shoreline configuration.

Brief remarks on the retreat of the western and eastern mainland shores

In the cliffs of the Bankeryd area and to the north of it the retreat since the latter part of the 18th century has been less than can be determined by map comparisons. The only exception is at the outlet of the river Dumneån, where a former road bank has been abraded, and the retreat can be estimated at about 12 m.

In isolated localities north of the Svedudden promontory a recent retreat of 5 to 10 m during the last 100 years can be estimated from field observations. Because there are forests in this area and no fields, the retreat cannot be determined by map comparisons. In the Guldkroken area there are also isolated erosion scars in the low moraine cliff but no general retreat of importance. Along the shore of the large estate of Almnäs there is an old, partly demolished, rock and timber protection. North of Hjo, there are only minor scars in the moraine cliff. The only area with an active retreat is the moss shore in the bay 9 km south of Karlsborg.

On the eastern shore immediately north of the Huskvarna bay the retreat since 1794 (E. 52 1^{4}) can be estimated to have been about 15 m (cf. Weiler 1926, Fig. 3: "shoreline 1799"). This shore is now stabilized by boulders. Further north the recent retreat has been very small.

In the Gränna area the retreat within the municipal boundaries between 1805 (E. 42 1^{3}) and 1858 was, according to Allvin (1859, p. 24), about 10 m (34 Swed. ft), or annually about 0.20 m. Since then this can be regarded as being the rate of maximum retreat in the Gränna area, and the average retreat is estimated to be less than 0.10 m/year.

The most northerly eastern shore with a recent retreat of importance is that of the moraine cliffs along the Ostrogothian Plain. No detailed map comparisons were made in this area, but from comparative studies of the morphology of the cliffs the maximum retreat of the most exposed parts north of Mount Omberg can be regarded as about 0.2 m/year. However, judging from the vegetation and the number of boulders, the main part of this shore is estimated to have been almost stable for at least 100 years. From the area north of Vadstena Johansson (1955) has records of a maximum retreat of 11 m during the last 25 years and 5.5 m during the last 8 years. However, even in this area the predominant part is of a stable character, and thus the average retreat is insignificant.

Conclusion

The recent shore retreat, as estimated by map comparisons and observations, corresponds to the morphological distribution of the shore types. The retreating shores, distinguished by their cliff morphology, are found in the area of submergence, and the high rates of retreat in the southern part of the lake. But the distribution is also closely connected with the kind of the deposits and the exposure to wave action. The most rapid retreat is in areas where there is not enough residual material, and where the abraded masses are continuously lost by beach-drifting.

As could be expected from the distribution of the submergence, the greatest contrasts are found at the southern end of the lake, where, on the one hand, there is a rapid retreat of the exposed Rosenlund cliffs, although the cohesive sediments are rather resistant to abrasion and, on the other, there is an accretion, in spite of the submergence, in the surrounding bays by beach-drifting.

INVESTIGATIONS ON THE BOTTOM MORPHOLOGY

General sediment distribution

Introduction

The general trends of the bottom topography have been presented in the chapter on the geology of the Vättern basin. In this chapter the sediments and submerged fossil morphology will be dealt with. The organic sediments are only briefly discussed, partly because they have previously been investigated by Ekman (1914, 1915), and partly because they involve problems beyond the aspects of this paper.

The bottom sediments are of great interest from the point of Quaternary geology. *A priori,* they could be expected to reflect the conditions during the stage of ice recession and the following stages controlled by Post-Glacial uplift. A description of the presumed conditions as regards the basin in general will be found in the introductory section on the Quaternary geology. In this study the aim has been to analyse the character of the bottom sediments and the fundamental features of their distribution, but the study is not concentrated on the geochronological problems.

From the start of the investigations it was clear that it might be problematic to distinguish between sediments deposited in very recent times and in times, when in the southern parts of the lake the water level was far below the present level. Thus different kinds of traces of previous conditions had to be searched for in different areas.

The further southwards the lowest Post-Glacial shoreline could be identified the better, because this would give the most accurate value of the slope of this shore in relation to the present water level. However, around the southern end of the lake there was considerable erosion during the stage of submergence, and the lowest shoreline could be expected to be hidden by sediments deposited during this stage. Another problem is that the shoreline expected to lie at a depth of about 40 m could not be effectively examined by divings.

The lowest shoreline intersects the present water level at about the latitude of Motala. In the area nearest to the south of this intersection recent wave activity affects the same depth interval as that of the lowest stage, and thus superficial sediments of different ages cannot be distinguished. The area most suitable for investigation with regard to the lowest shoreline could be estimated to lie between the latitude of the Flisen shoal and the latitude of Gränna. Between these points the expected depth of this shoreline increases from about 13 m to 31 m (cf. Fig. 7).

Because the submergence was most rapid in the southern part of the lake, relicts of relatively recent submerged shores could be expected to be found in that area.

Fine glacifluvial sediments should be looked for in the even central part of the plateau north of Visingsö, that lies below the lowest shoreline. There the possibility of finding a thick cover of Post-Glacial and recent sediments could be expected to be small because of the distances to the present shores. If varved clay was found in this area, open water must have existed in the whole central and southern part of Vättern during the final ice recession.

Glacifluvial sediments of sand and gravel could primarily be expected in the northern section of the lake, where large glacifluvial deltas are found in the surrounding lands.

In the following sections of this chapter previous investigations are first reviewed, and then the distribution of the superficial bottom sediments found in them is discussed. These sections outline the general distributional pattern. In the subsequent sections the series of fine sediments expected from Glacial to recent time is presented and discussed on the basis of cores sampled in the central part of the lake and investigations on the bottom east of Stora Röknen island. The chapter concludes with sections dealing with direct investigations on the lowest shoreline in the central part of the lake and on higher submerged shores of the southern part.

Previous investigations on the bottom sediment

The first information about the bottom sediments of Lake Vättern was obtained from the hydrographic survey of 1816–17 (the Hydrographic Office, reg. no. XX). The bottom was examined with the usual surveying lead with a cavity in its foot. As always in charts, the description of the sediments is very restricted. In this first chart, as well as the chart of 1879 still current (Chart no. 12, formerly no. 81), a remarkable distribution of the bottom sediments is found. Sand and also gravel are noted at great depths and far from the shores in areas where only fine sediments can be expected from the present conditions.

The first scientific investigation on the bottom sediments was carried out by Ekman. This investigation was primarily directed to the bottom fauna (1915), and the inorganic sediments and their origin were discussed in a special paper merely as a by-product (1914). Ekman sampled at most the upper 7 cm of the sediment with a grab sampler invented by himself (1911). No quantitative grain-size analyses were made. On the whole his distribution is consistent with that of the chart. The sand was often found to be mixed with finer particles, or there was only a very thin sand layer that covered fine sediments. In one case a layer of 5 cm of pure sand was found 1000 m from the shore. A sediment type described as "more or less pure clay" was found less often than sand as regards the actual surface. This type was sampled from depths between 9 and 88 m. In four cases the sample was interpreted as Glacial varved clay. Mainly organic mud (gyttja) was only found in the deep troughs.

In Ekman's map on the bottom sediments (1914, Fig. 3) four types of sediments are distinguished: sand or sandy sediments, pure clay, kitin gyttja, and fine gyttja. It should be pointed that, in Ekman's terminology, sand probably includes coarse silt, and clay includes fine silt. The sand or sandy sediments are shown as covering the whole bottom, except the deep troughs with their gyttja. Pure clay is noted at six points, at the margin of the troughs or within them. The kitin gyttja is an organic mud characterized by preserved fragments of crustacea and of oligochaeta excrements. In the fine gyttja their structure has been destroyed to a greater extent.

At the time of Ekman's investigation little was known about the Glacial and Post-Glacial water levels of Lake Vättern, and he tried to interpret the sedimentary distribution from the present hydrographic conditions. The sand is interpreted as a product of Glacial (varved) clay, eroded by assumed circulation currents still present. This process is said to have ceased as soon as a thin covering layer of sand had been formed (op. cit., p. 362).

Lundqvist (1942 *a*) discusses the Vättern sediments as described by Ekman from the point of view summarized by the question: "Where are the ice-lake sediments?" He points out that a Glacial clay does not have a composition which can result in a cover of sand. His conclusion is that the clay is not a sedimentary varved clay but clayey moraine, and in a map (1942 *b*) the whole bottom, except the deep troughs, is marked as consisting of clayey moraine. Lundqvist's interpretation was supported by a sample from a core taken by Kullenberg between Hjo and Hästholmen at a depth of 43 m. Its analysis showed 50 % clay and 50 % poorly sorted material, ranging in size up to granules. This distribution is said to be like the clayey moraine of an oscillation zone, where moraine has been mixed with sedimentary clay. Lundqvist agrees with Ekman's idea of circulation currents in the lake, which have an erosional effect, and he accepts the sand as a "residuum".

Professor Kullenberg has informed me that the core referred to by Lundqvist was the only one taken. Through the boat's drifting the core sampler was broken in the first attempt, and the material given to Lundqvist came from the broken stump that was saved.

If practically the whole of Lake Vättern has a surface deposit of moraine, there must have been very special conditions during the ice recession. "The consequence is that the Vättern basin was not open during the recession but filled with a body of ice" (Lundqvist 1942 *a*, p. 161). However, this explanation conflicts with the map later published by him (1962), in which there is assumed to be open water south of the retreating ice margin.

Several objections can be raised against these explanations of the origin of the sand cover. It certainly seems more probable that the sand emanates from moraine than from varved clay. However, a clayey moraine is very resistant to

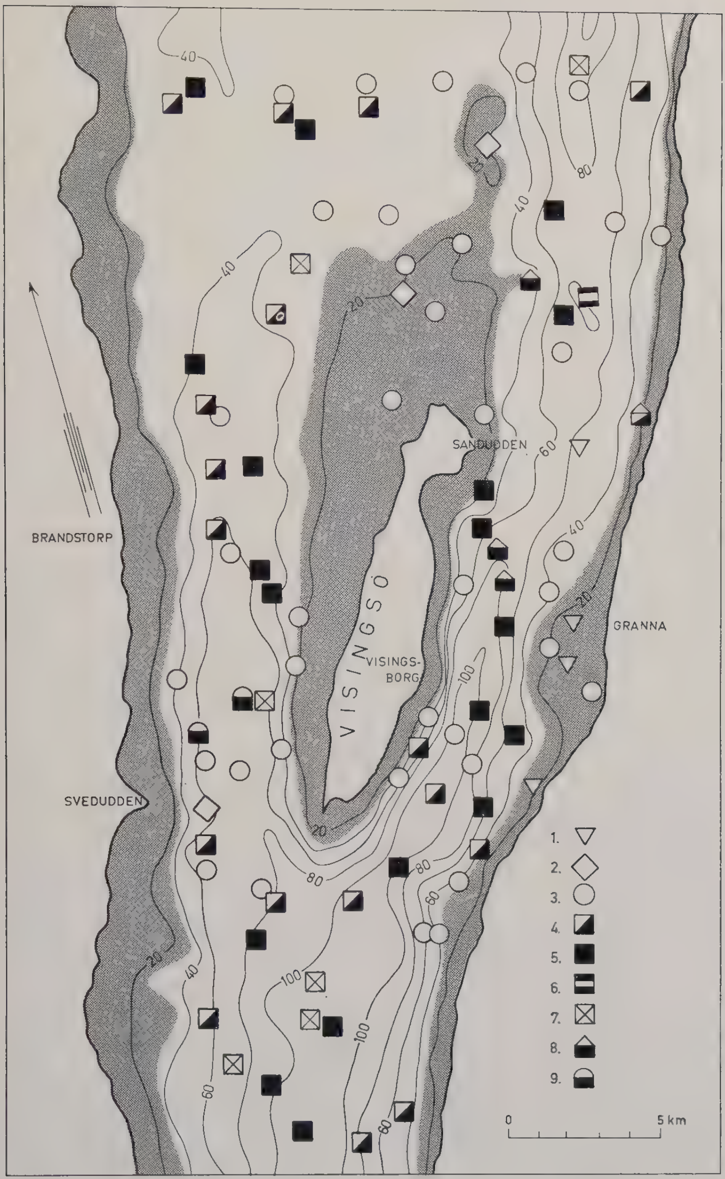


Fig. 126. The sediment distribution of the central section around Visingsö island according to the Chart and Ekman's samples. Areas above the lowest shoreline predicted by E. Nilsson (1953) are shaded. 1, cobbles. 2, gravel. 3, sand. 4, fine sand–clay. 5, clay. 6, varved clay 7, gyttja. 8, gravel and sand on clay 9, sand and silt on day.

erosion. There are no circulation currents, which can erode the moraine and spread sand over the whole bottom (cf. p. 64), and a thin cover of sand is not a "residuum" in relation to a clayey moraine.

The previous discussions on the Vättern sediments have not solved the problem of the character of the Glacial deposits, nor have the Post-Glacial effects been satisfactorily explained.

The superficial sediment distribution in the central and northern sections

If the low Post-Glacial water level predicted by E. Nilsson (cf. Fig. 7) by extrapolation of shorelines outside the area of present submergence really existed, it should have considerably affected the distribution of the bottom sediments in the central and southern parts of the lake. Thus the position of the lowest

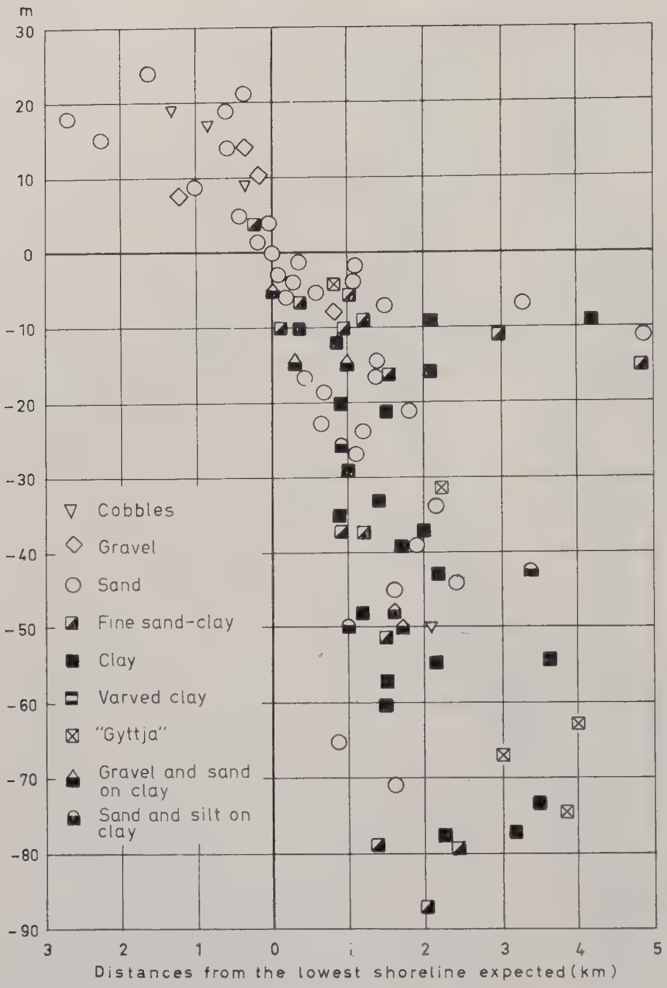


Fig. 127. The sediment distribution of the central section around Visingsö island in relation to levels and distances to the lowest shoreline predicted by E. Nilsson.

shoreline can be tested by the sediment distribution, and if the predicted level is verified, this will also explain the widely spread sand cover.

In the northern section (north of Motala) the water level has never been much lower than at the present time. Thus the sediment distribution in that area should differ from that in the central and southern parts of the lake.

Professor Ekman was kind enough to let me have his original field notes from the investigations in 1911 and 1912. In these notes the following grain sizes are distinguished: stones (in a few cases "small stones"), sand (in a few cases "fine sand") and clay. The clay is characterized with regard to colour, consistency and stratification (varved, strata of sand, homogeneous). The occurrence of coarser particles mixed with the clay is also noted (sand grains, stones). The colour of the clay varies considerably: brownish-red, yellowish-grey, bluish-grey, grey. The reddish-brown clay is the most common. This colour entirely predominates in clays mixed with sand. The consistency varies from "hard cakes" to "sludge". The large jumps in the grain-size classification are problematic. From the description in Ekman's notes it is obvious that "stones" as a rule should be called gravel—"stones up to 2 cm in size". As previously mentioned, "sand" probably includes coarse silt.

In Fig. 126 the position of Ekman's samples in the central section, supplemented by the bottom characteristics according to the chart, are shown. The areas shoreward of the intersection between the lowest Post-Glacial water level, according to E. Nilsson (1953), and the bottom topography is shaded. The lower limit of these areas does not show the exact lateral position of the lowest shoreline, because it is based on the present topography.

If the samples are referred to the present shoreline and water level, the distribution with regard to grain size will be found to be very unsystematic. This fact was noticed by Ekman (1914, p. 361), who believed it to be the general condition of the whole lake.

In Fig. 127 the position of the samples of Fig. 126 is shown in relation to the lowest shoreline expected with regard to vertical and horizontal distances from this shore. The distribution of the points as such reflects the topography of the area with its deep channels and shallow plateau. The points, which lie at a depth of about 10 m and far from the shore, belong to the area north of Visingsö. In general the distribution of the different types of bottom sediments is immediately explained by this diagram. Above the lowest shoreline cobbles, gravel and sand entirely predominate. These sediments have been formed in the shore zone during the time of the sinking water level and the following submergence. As could be expected, sand also predominates down to about 7 m below the lowest shoreline. Below this level there is an increasing amount of fine sand (silt) and clay. However, anomalies are found. There is a cluster of points with "gravel and sand on clay" and cobbles. These samples were taken in the northern part of the channel east of Visingsö. The sample, which has been marked as gravel, was mixed with clay. Ekman noted: "Yellowish-grey clay with stones, the largest one 5 cm". This may be a clayey moraine. As regards the few samples with sand and/or gravel on clay, the coarse material may have originated from floating ice, either from glacial ice or from lake ice. If there was open water south of the glacial-ice margin, icebergs can be expected to have dropped material on the bottom. In the spring of 1956 it was observed how shore ice with strata of frozen sand and gravel drifted from the shore. Sand and even coarser material can also have been carried into the deep channels by gravity movements. Off Sandudden on Visingsö sand deposited under the ultimate angle of repose has been examined on the steep slope facing the channel.

It can be stated that in the central section the sediment distribution—except for a few anomalies—can be explained as having been formed by normal shore processes, and that the lowest Post-Glacial water level predicted by E. Nilsson agrees well with the distribution.

The same type of diagram, based on samples from the open parts of the northern section, where the water level has never been lower than at the present time (north of lat. N. 58° 35′), is shown in Fig. 128. In this diagram the positions are given in relation to the present shores. The main parts of the shores of this area are rocky or consist of wave-washed, rather blocky moraine, but there are also some sandy bays. No nearshore samples are included in the material presented in Fig. 128.

At depths greater than 20 m only 4 samples out of 36 consist of sand. The point marked as "cobbles" is taken from the chart, and there it is marked as "stones or rock". The locality is found off the skerries of the Tiveden Forest, where the bottom is very irregular with vertical walls and narrow shelves. The four sand points marked by arrows have an anomalous position. They have also been taken from the chart. The one at a depth of 35 m is found on the platform 10 km north-east of Karlsborg (cf. Fig. 5). Because of its depth it cannot have been directly affected by waves, and it is surrounded by deep channels, which rules out the possibility of transportation from a shore. Thus the sand must be interpreted as being of glacifluvial origin. The point more than 8 km from the shore is found at the outer edge of the shallow platform SSE. of Stora Röknen. The sample of sand mixed with clay from a depth of 87 m was taken at the foot of the same slope. The point at a depth of 70 m lies in the narrow northern end of the deep channel west of Stora Röknen. Finally, the sample from a depth of 18 m was taken on the isolated ridge east of Stora Röknen and, because this ridge is surrounded by deep channels, the sand cannot have emanated from the shores. This locality is dealt with in the special section on the bottom conditions east of Stora Röknen.

In the surroundings of the northern section of the lake there are several glacifluvial deltas

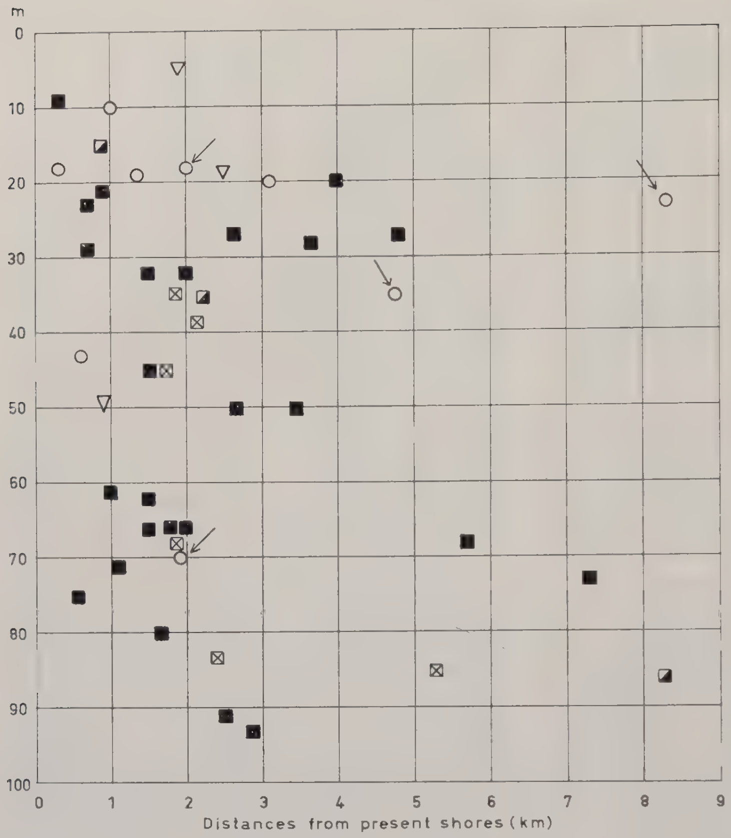


Fig. 128. The sediment distribution in the northern section of the lake with regard to depths and distances to the present shores. Data according to the Chart and Ekman's samples.

and eskers, and it seems reasonable that some glacifluvial sand should also be found within the lake. But for the above-mentioned anomalies in relation to a sediment distribution expected from shore processes, the diagram does not give rise to any extraordinary interpretations. The clay that predominates in the diagram may be of Glacial origin, or it may be a cover of suspended material from shore processes.

Hypothetical series of fine sediments

The previous investigations only referred to the composition of the superficial bottom sediments, but from our knowledge of the deposits around the lake and the Quaternary development a hypothetical series of fine sediments can be sketched. This series has never been complete within the whole lake region, and in large areas it is partly or totally abraded.

The outline of Table 21 refers to the plateau levels below the lowest shoreline in the area north of Visingsö. It is based on the development described in the introductory chapter on the Quaternary geology, and it is assumed that there was open water in this area before a period of ice-front oscillation, and that again there was open water during the final ice recession. The presence of an oscillation is documented by the varved clay that intercalates the moraine of Visingsö.

The relation between the fine material emanating directly from the ice and the material washed out from the shores south of the ice margin is uncertain. It depends on the subglacial drainage pattern and the distances to the shores. Areas which by the rapid lowering of the level of ice-dammed water came into a state in which material from the shores could be deposited can be expected to show discordances in the sediments (Sauramo 1924, p. 105). In the Vättern basin such areas were generally passed by the shoreline during the

Table 21. Hypothetical series of sediments from central Lake Vättern.

Sediment	Water	Water level	Sources	Sediment character
Vättern sediment	Fresh	Rising	Shore processes	Decreasing grain size, non-stratified or with organic layers. Gyttja in the troughs.
Yoldia sediment	Salt (brackish)	Falling	Shore processes	Increasing grain size, coagulation of clay, non-stratified or with organic layers. Gyttja in the troughs.
Glacial Yoldia clay	Brackish	Falling	Shore processes and glacifluvial flow	Decreasing grain size, coagulation of clay, symmict stratigraphy.
Baltic Ice Lake clay	Fresh	Falling	Glacifluvial flow and shore processes	Decreasing grain size, varved clay.
Vättern Ice Lake clay	Fresh	Falling	Glacifluvial flow (shore processes?)	Decreasing grain size, varved clay (homogeneous clay?).
Moraine	(Fresh)	?	Glacier ice	Clayey moraine with embedded glacifluvial sediments.
Vättern Ice Lake clay	Fresh	Falling	Glacifluvial flow	Varved clay (homogeneous clay?).
Moraine	—	—	Glacier ice	Clayey moraine.

Yoldia stage, and the sediments have been abraded.

The extent to which sediments of the different stages of Table 21 can be distinguished by stratigraphic differences is difficult to estimate. The main factors which could yield specific conditions are, on the one hand, the genetical origin—primary glacifluvial deposition and re-deposited material from the shores—and, on the other, the depositional conditions—fresh and salt (brackish) water and different current patterns due to change of the position of the outlet.

Glacial varved clays have principally been used as a tool in geochronological studies. In this way the false idea of an identity between fine glacifluvial ice-lake sediments and varved clay may arise. On the contrary, it has been found difficult to state the relations between depositional conditions and stratigraphy, grain-size variation and colour. The geochronological method based on the varves presumes that the proximal varves (bottom varves) can be identified. However, in investigations on the Glacial clay of the Rönneån river in Scania Tullström (1955) found as a rule that a homogeneous or stratified clay without intra-grading formed the lowest part of the series. The absence of a varved structure is explained as having been caused by a high concentration of suspended material nearest to the ice margin. The same interpretation has been given by Hansen (1940), as regards Danish and Scanian clays. The sediments described by Tullström are usually grey, but some examples of brownish clays are mentioned.

Drillings performed by the Geological Survey in the valley of the Göta Älv river have in several cases proved the existence of brownish-coloured proximal varves. According to Tullström (1961, p. 16), this colour can be explained by the presence of iron compounds of the glacifluvial material already oxidized in the bedrock source and not reduced because of lack of organic substances. According to Tamm (1925), the combination of a high concentration of suspended material and a high, partial, carbon-dioxide pressure in the subglacial meltwater induce a tendency to coagulation and an oxidization of the sediment.

The general opinion is that there is a grading within the yearly varves with the coarsest material deposited during the meltwater flood in late spring and summer and a decrease in grain size during the diminishing flow of the autumn. A coincident change of the tint from light to dark could thus be explained as a function of the grain size. However, in an analysis of some calciferous varves from Uppsala the same seasonal variation of tint was found, but there was no co-variation with the grain size (Arrhenius 1948). The seasonal variation could be explained by a higher percentage of organic material (plankton), settled during a season with a small supply of inorganic particles. The importance of the organic content as regards the tint variation is generally accepted (Lundqvist 1957), but it should be stressed that in most varved sediments there is also a seasonal variation of grain size.

In his great work on the stratigraphy of the varved clays in Finland Sauramo (1924) to a large extent avoids characterizing the strata by colour, but in his descriptions he distinguishes different dark and light shades.

The discordance within the fine sediments of the Vättern basin that could be expected to be most pronounced is the one connected with the transition from fresh water to brackish or salt-water conditions. Sauramo (op. cit) explicitly discusses how, by coagulation in the brackish or salt water of the Yoldia Sea, aggregates of clay settled together with coarser single particles of the same settling velocity as the aggregates. In this way symmict strata were deposited in contrast to the diatactic varves of the Baltic fresh-water stage. Thus the Yoldia clays could not be expected to have well-sorted unimodal grain-size distributions. Judging from Sauramo's descriptions and figures, it should be possible to distinguish the symmict strata ocularly. These strata are generally thicker than the subjacent diatactic varves, and the symmict series has a banded appearance.

In the area around Stockholm the transition from Baltic Ice-Lake clay to Yoldia clay is characterized by grey, relatively coarse-grained varves, abruptly succeeded by thick brown strata (De Geer 1932, p. 26). If sea water could penetrate into the Vättern basin, when a glacifluvial suspension was still flowing from an ice margin in northernmost Vättern, a similar transition could be expected. In large areas a rapid penetration of sea water ought to have caused a sudden predominance of sediments abraded on the shores, because by coagulation the main part of the fine glacifluvial settled near the ice margin. The transportation of suspended material from the shores can be expected to have been less affected by the brackish water, because there the concentration of solid particles was probably much lower than in the subglacial meltwater. However, the connections with the sea in the Karlsborg area were narrow. If the Baltic Ice Lake was first lowered 25 m, and then all drainage from the Baltic had to pass this way, these connections would mainly have the character of outlets. Thus it is uncertain if the water became significantly brackish before the ice had left the northern part of the basin.

At the end of the Yoldia stage a grain-size maximum in material from the shores is expected in areas of continuous sedimentation. At this time the water level in the central and southern parts of the lake reached its lowest point. During the following submergence with transition to fresh water and growing distances to the shores the grain size is supposed to have decreased.

Bottom cores from the central section north of Visingsö island

Description

Seven cores sampled in 1961 and 1962 are discussed. The position of the sampling points (*a*–*g*) is shown in Pl. 1 and the table below.

Table 22. Position of cores sampled in the central section.

Core	Position	Depth (m)	Depth below lowest shoreline
a	N. 58° 19.3′ E. 14° 24.5′	28	12.5
b	N. 58° 17.8′ E. 14° 23.8′	38	21.5
c	N. 58° 17.1′ E. 14° 24.5′	38	20.5
d	N 58° 16.7′ E. 14° 28.9′	44	26.0
e	N. 58° 08.3′ E. 14° 17.3′	43	18.0
f	N. 58° 14.1′ E. 14° 28.7′	38	17.5
g	N. 58° 16.2′ E. 14° 31.0′	60	42.0

The main characteristics of the stratigraphy of the cores are illustrated in Fig. 129. The cores *a*–*e* consist of grey varved clay in the lower part and of brownish clay in the upper one, where also the varve structure disappears. In the cores *a*, *b* and *e* there is sand and silt at the top. The cores *f* and *g* are all brownish and without any varve structure. In core *g* there are some diffuse variations of light and dark shades.

There is one certain discordance in the cores, that between the varved sediment and the ocularly homogeneous brown clay. If the sand and silt at the top of cores *a*, *b* and *e* is accepted as having been washed out from the shores at the time when the water level was lowest, this clay was deposited during the Yoldia stage. The opinion that the sandy sediments have come from the shores is supported by the fact that the amount decreases from *a* to *b*, and that there is no sand in *c* and *d*. The

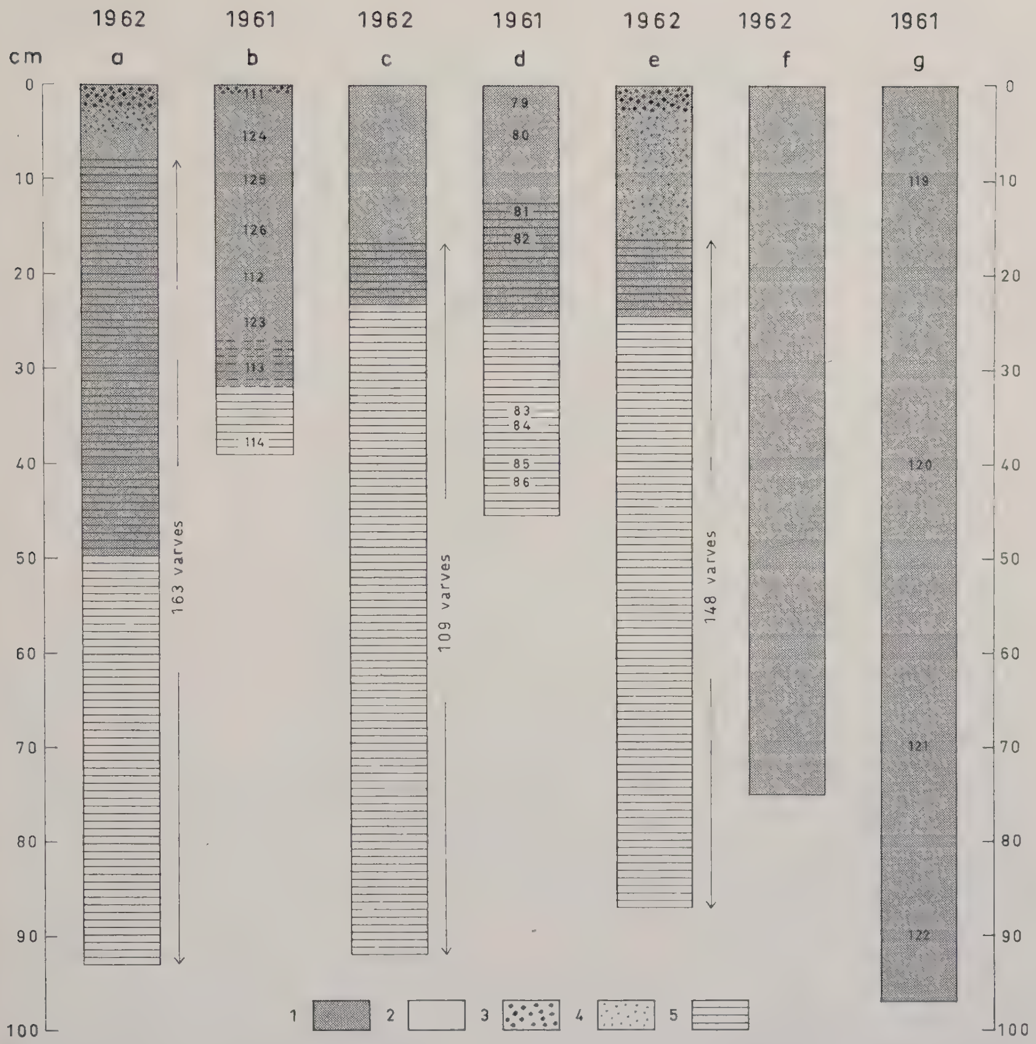

1962
1961
1962
1961
1962
1962
1961
cm
a
b
c
d
e
f
g
0
10
20
30
40
50
60
70
80
90
100
111
124
125
126
112
123
113
114
79
80
81
82
83
84
85
86
119
120
121
122
163 varves
109 varves
148 varves
1
2
3
4
5

Fig. 129. General stratigraphic characteristics of cores from the central section of Lake Vättern. (1) brown, (2) grey, (3) sand, (4) coarse silt, (5) varved clay.

thick deposit in core *e* can be explained by its nearness to the shore.

The stratigraphy of the varves has not been studied with respect to geochronological problems. In the varve series of core *a* 163 varves have been identified, in core *c* 109, and in core *e* 148. The thickness of the varves varies from about 1 to 0.2 cm. This sediment was obviously deposited in a distal position in relation to the ice margin.

The grain-size distribution of cores *b, d* and *g* has been analysed. In core *b* there is only 7 cm of varved clay. The darker parts of the varves have been shaded in Fig. 130. The dashed lines mark very thin (1.0—0.5 mm) dark layers in the lowest part of the brown clay. The sandy sediment at the top has a thickness of 2 to 3 mm. Throughout the core there are rust-coloured stains with a size of about 1 mm, and on an average one per cm^2. They are probably caused by oxidization around some organic substance. Eight analyses of the grain size have been made. From the varved clay the light parts have been prepared. In the diagram of Fig. 130 values of median, quartiles, 90th and 10th percentiles are shown in relation to

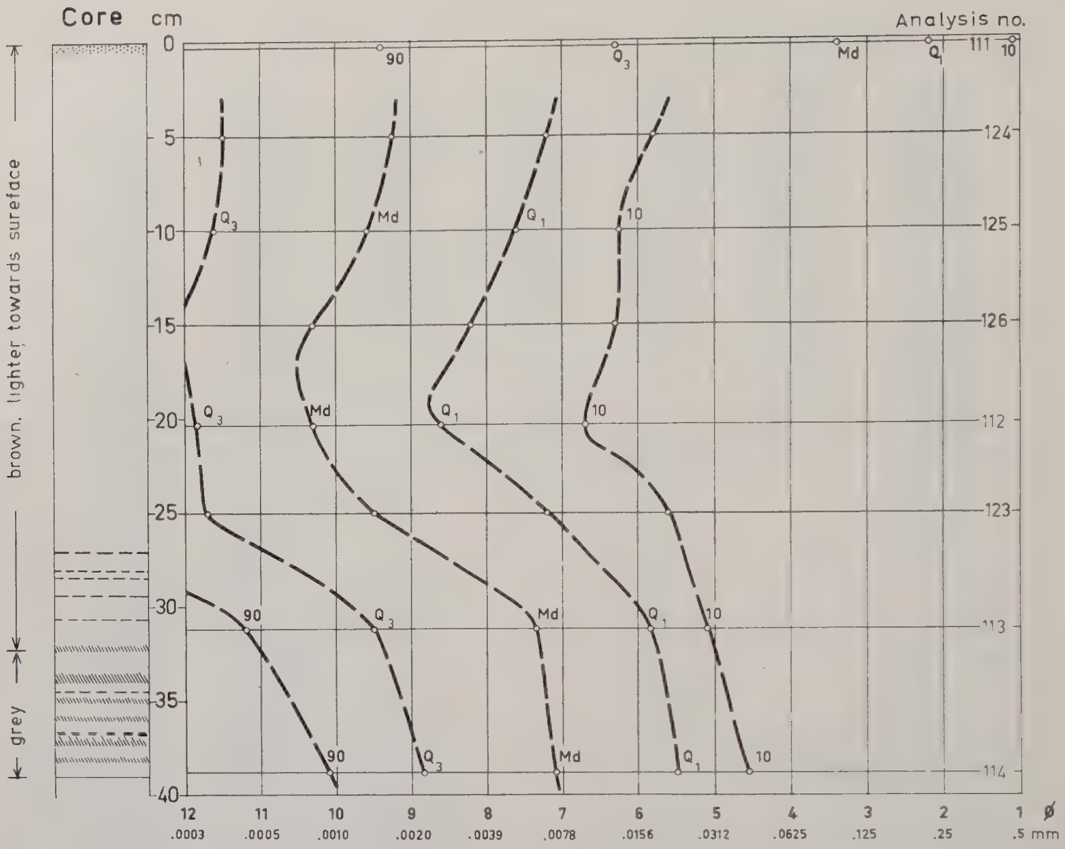


Fig. 130. Core *b* from the central section. Dark strata of varved clay shaded. Very thin dark strata marked by hatched lines. Grain-size distribution according to eight analyses (1961).

the positions of the samples. The variation with level can be judged to be systematic. At the transition from the varved part to the brown one the size of the particles drops; the median falls from about 7.5 Ø to about 10.5 Ø at 15–20 cm below the surface. Then there is a slight increase. As regards the quartiles, the dispersion is much the same. Because the sandy sediment at the top is so thin, the analysis (no. 111/61) contains both this sand and finer material. The distribution is bimodal, but, if it is separated, one finds that the surface material mainly consists of medium and fine sand (0.6 to 0.06 mm).

In core *d* the grey part has a length of 21.5 cm, and the varves have an average thickness of 0.9 cm. In the brownish-grey part above that (11.5 cm) the varves are thinner; in the lower part they are about 0.6 cm and in the upper part 0.5 cm. Four analyses have been made of the grey part, two of the brownish-grey and two of the brown unstatified clay (cf. Fig. 131).

In the grey clay the dark part of the varves was separated from the light one. The light part of the upper varve analysed (analysis no. 83/61) is significantly coarser than the dark part (84/61), but in the pair 85 and 86/61 there is no significant difference. It can be observed that the medians of the latter varve lie between the medians of the former. The distributions of the latter also have a tendency to bimodality with a minimum in the interval 5–6 Ø at the 2–3 % level. Therefore it seems probable that the separation of the lower varve failed.

The varves of the brownish-grey part were too thin to be separated. From this part to the brown one the grain size of the median decreases about 1 Ø unit. The grain size of the

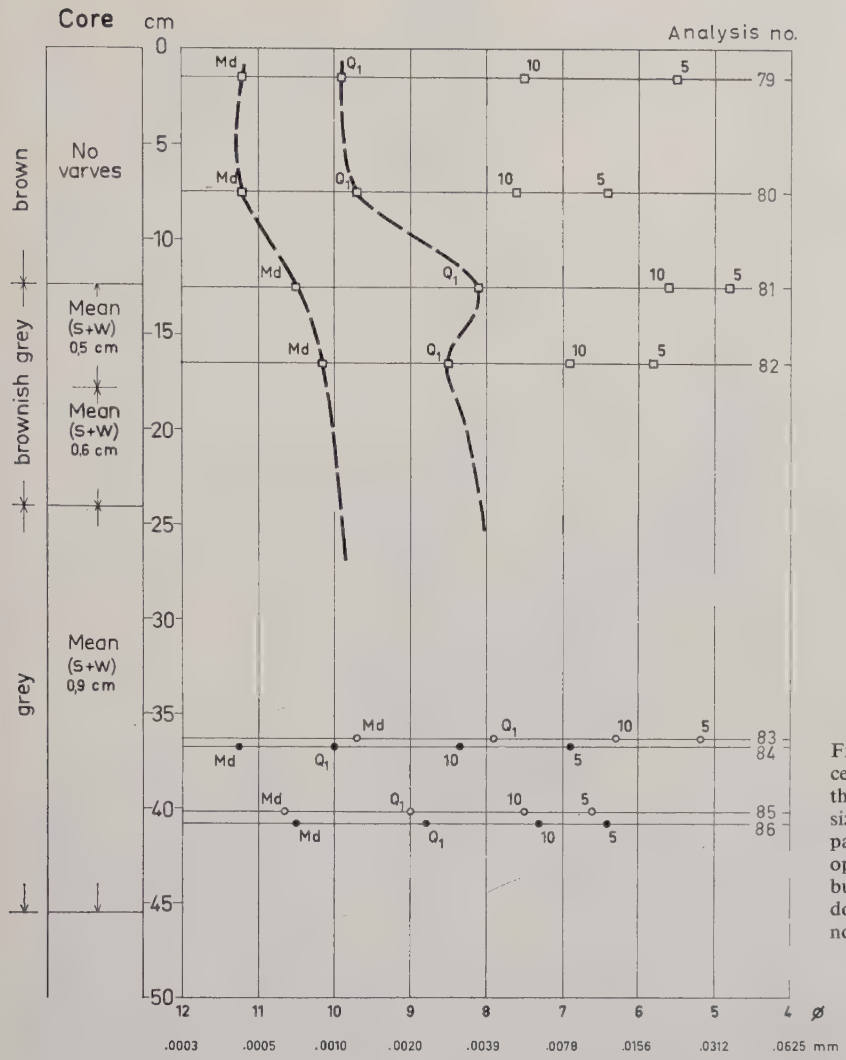


Fig. 131. Core *d* from the central section. Mean thickness of varves. Grain-size distributions of light parts of varves shown by open circles, and distributions of dark parts by dots. Open squares mark non-separated varves.

brown part is very fine; the median of analysis no. 80/61 is 11.2 Ø, but, as can be seen from the diagram, there is also a small proportion of rather coarse material; 5 % is coarser than 6.4 Ø in the same analysis and coarser than 5.5 Ø in analysis no. 79/61. The top of this core is comparable with the –15 cm level of core *b*.

The grain-size distribution of core *g* has been analysed at four levels: –10, –40, –70 and –90 cm (analyses nos. 119–122/61). The distributions are all of the same type, but the two lowest samples have a greater skewness because of a larger proportion of material coarser than very fine sand. The relation between the distributions can be illustrated by a table.

Table 23. Percentages of material in different grain-size intervals of core *g*.

Level cm	Grain-size intervals (μ)		
	< 2.0	2.0—15.6	> 15.6
—10	38.90	35.05	26.05
—40	48.73	35.23	16.04
—70	12.22	26.08	61.70
—90	28.58	27 61	43.81

The proportion of fine silt and coarse clay is of the same order in all the samples; about one-third in the upper ones and about 10 % less in the lower ones. The differences in the coarser material are more pronounced. This sediment can be estimated to consist of a fine silt-clay suspension having a normal logarithmic distribution with small variations as regards dispersion, and a variable proportion of coarser material.

Discussion

The find of varved clay at five points out of seven, and the fact that no till was found, is, at least in part, an answer to the question: "Where are the ice-lake sediments?" The core sampled by Kullenberg must be regarded as an unlucky strike on a "morainic float" (a deposit of moraine dropped from an iceberg).

The varved clay of the cores was obviously deposited in a distal position in relation to the ice margin. The presence of varved clay in the central section of the lake means that there was open water south of the ice margin during the final ice recession. It seems probable that further knowledge of the ice recession can be obtained by the use of a larger piston core sampler.

The exact origin of the brownish unstratified clay is uncertain. It is found in all the cores, but it is much thicker in cores *f* and *g*. The amount could be expected to increase with depth, but core *f* is not from a depth greater than the other ones from the plateau. However, cores *f* and *g* have a different position. Core *f* was sampled at the margin of the plateau facing the deep eastern channel, and core *g* was sampled on the slope of the same channel.

During and after the lowering of the level to that of the Yoldia Sea the depositional conditions of the basin changed in many respects. Before the lowering the water depth of the plateau was of the order of 85 to 100 m. Immediately afterwards it was about 22 to 25 m less. Before, the basin was only fed by meltwater from the basin itself and possibly also from the basin west of the Hökenås hills (the Tida basin), and the flow was towards the Baltic. Afterwards, suspended material from the whole ice margin of eastern Sweden had to pass through the basin during its westward flow, and it seems probable that in this flow the deep eastern channel of the Vättern basin served as a sediment trap. This model explains the thick deposits of unstratified clay on the slope and margin of the channel. The brownish sediment is somewhat finer than the light parts of the varved clay, but the grain-size distribution has not been proved to be an effect of brackish water. The cause of its colour is uncertain; it may have been caused by oxidization during a long transportation and by a higher content of organic material in the water.

From the general distribution of sand in the central section and from its distribution in the cores it seems certain that this sand was deposited during the lowest water stage of the basin. This means that there has been almost no sedimentation on the central parts of the plateau north of Visingsö during the last 9000 years. The absence of late inorganic sedimentation is due to the distances to the shores and the depths of the plateau. The absence of organic material can be explained by the poor organic production and the fact that plankton of small weight, in contrast to the mineral grains, can be carried by slow currents to the deepest parts of the basin.

The area east of Stora Röknen island

This area was specially investigated, because two isolated depth figures indicating shallow water (18 m) and a note of "sand" were observed on the chart, and these figures were surrounded to the west and the east by depths of 55 to 67 m. Thus the sand (if not erroneously mapped) could not have come from the present shores. It was of special interest to know the type of bottom sediments in this area, isolated from the shores, where, according to E. Nilsson (1953, Pl. IV), the depths have never been smaller than at the present time.

General bottom morphology

The area was echo-sounded in 1960 and 1961. The depth figures of the chart were found to agree with the echo-soundings, but the figures were too few to show the character of the morphology. A map of the area with 10 m contours was constructed from the new topographic map (1961), the echo-sounded sections, and the depth figures of Chart no. 12.

The bottom morphology is characterized by two channels with a depth of more than 50 m, which to the south enclose a central ridge (Fig. 132) that in the northern part reaches a height

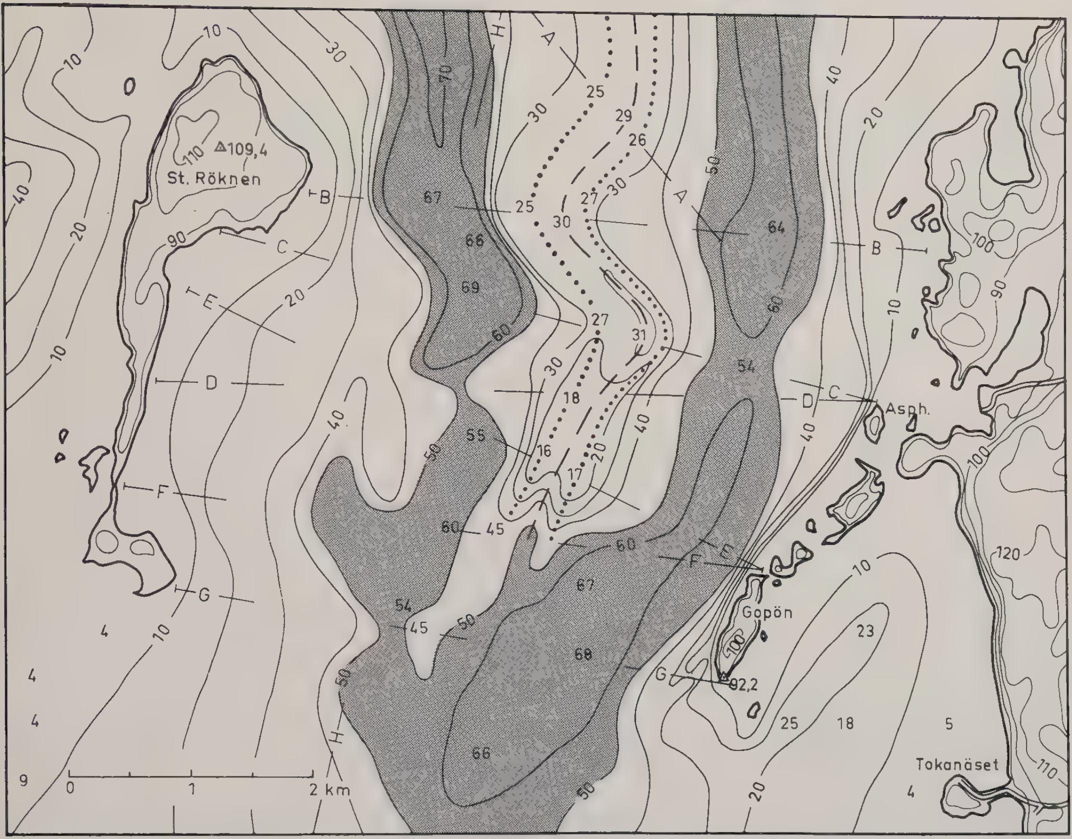


Fig. 132. Map of the area east of Stora Röknen island. Shore and land contours according to Fältkarta 9E Askersund SO. Bottom contours based on depths of Chart no. 12 and echo-sounded sections *A—H*. Dotted lines mark crests on the central ridge, dashed line marks channel between the crests.

of 25 m below the water level and near the southern end has a height of 16 m below the water level. (Through generalization this ridge is omitted in Ekman's (1914) and Lundqvist's (1942 *b*) bathymetric maps.) To the north the ridge seems to disappear, about 2500 m north of the mapped area, into a trough with depths greater than 90 m. The ridge runs east of the needle-shaped Lakaskär skerry (S in Pl. 1). The area north of that mapped in Fig. 132 has only been preliminarily surveyed by a few echo-soundings.

The slopes of the ridge are rather gentle. The contrast with the contours of the fault zone off Gopön island may be noted. The central ridge is either a narrow rib of rock bounded by faults or a glacifluvial accumulation of esker character. The transverse profiles have a very specific form, as can be seen from Fig. 133, which shows the upper part of the echo-sounded sections B to F. The ridge is double-crested, and the trench of the crest has a tendency to an asymmetry similar to that of river chanels. In section B the dashed contour is deeper in the western outer swing, in section C the same asymmetry is found in the eastern outer swing, and in section D the eastern crest is missing. In the rather straight course south of it (sections E and F) there is no specific asymmetry. It should be noted that the level of the crest differs considerably in the sections.

The origin of the ridge cannot be decided by the morphology, but it undoubtedly resembles the parts of the Uppsala esker in the Bothnian Sea, which have been echo-sounded (cf. Hoppe 1961, Fig. 3).

For the comparison between the sediments of an isolated ridge and the sediments of an

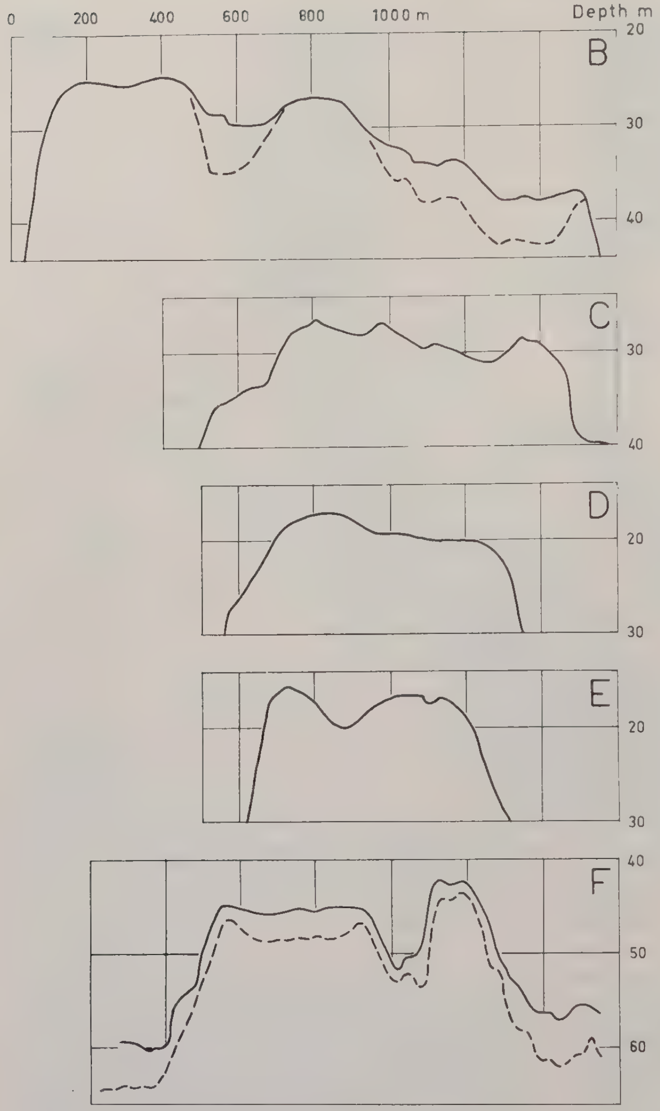


Fig. 133. Profiles of the upper part of the ridge east of Stora Röknen island. Dashed line marks discontinuity surface of the sediments indicated by the echograms. For position of the profiles, see sections *B—F* in Fig. 132.

area at the same latitude adjoining a shore, samples were taken off the shore of Stora Röknen island, and these samples will be discussed first.

The sediments off the eastern shore of Stora Röknen island

In the echo-sounded section C samples were taken at depths of 0.9, 5.1, 6.4, 7.7, 11.8, 16.5, and 26 m. Down to a depth of about 4 m the bottom is covered with sand (observations from diving operations). Further out the sand cover is not complete, and the bottom consists of a fine sediment with scattered boulders and cobbles. These coarse pieces lie on the bottom with only a small part of their volume below it. The exact outer limit of this material could not be determined, but the zone seems to be rather narrow.

The grain-size distribution of the samples (grab samples) are illustrated in Fig. 134. Nearest to the shore (d = 0.9 m) the sand (1–3 Ø) is extremely well sorted. The sample represents the most fine-grained well-sorted distribution

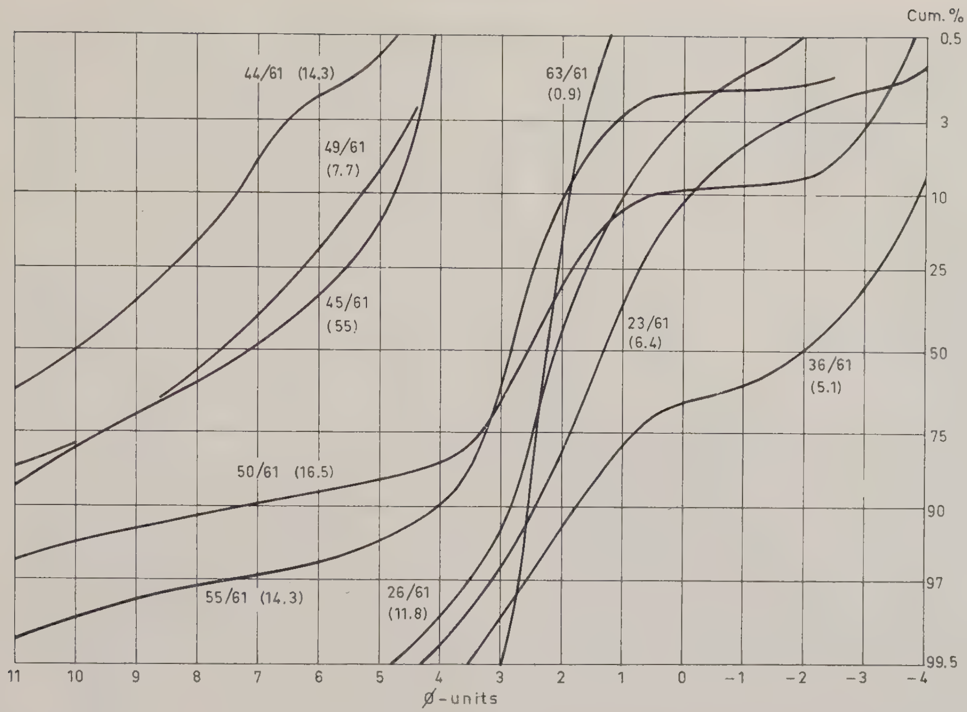


Fig. 134. Grain-size distributions of samples taken off the eastern shore of Stora Röknen island. Depths in metres in parenthesis.

that can be expected in a true bed load. The samples from 6.4 and 11.8 m have distributions with about the same mode as the previous one, but the sorting is not so good.

The analysis (36/61) of the sample from 5.1 m shows a bimodal distribution, consisting of one maximum in the sand fraction and another in the gravel fraction. There was also a pebble in the sample, not included in the percentual distribution, with a weight of 100 g. The total weight of the dried sample was 577 g. The part coarser than 5 mm (–2.3 Ø) consisted as to 11.6 % of iron precipitates (*ärtmalm* [pea ore]).

The sample from 7.7 m only consists of fine particles and iron precipitates, which were excluded from the analysis (49/61). The distribution of this sample is much like the distribution of a sample from a core taken at a depth of 55 (45/61, cf. Fig. 134).

There is finally the sample from 26 m, which is a core sample with a length of 57 cm (Fig. 135). Thirteen grain-size analyses were made on this core. The upper 16 cm has a brownish colour, which shifts from yellowish to brownish tints. Down to –5 cm there are small (0.1 mm) brownish-black precipitate stains, and at the –10 cm level a zone of reddish-brown spots. Between –16.5 and –18 cm there is a rather hard yellow-ochre stratum. Below that the colour is yellowish-grey down to –21 cm, where there is again an ochre-coloured stratum, but this one is soft. The lower part of the sample is light grey without any ocular traces of stratification.

Five analyses were made of the lower, grey part. All these show the same type of grain-size distribution. There is no consistent change of the grain size with the level. The median oscillates about the upper limit of the clay fraction (2 μ). Six analyses were made from the transition to the lower ochre-coloured stratum to the spot zone at the –10 cm level. The analyses show a consistent increase in grain size towards the surface. The median grain size increases from 9.3 to 5 Ø. Two analyses from the –5 and –1 cm levels show that in the uppermost part the grain size again decreases.

The upper limit of the light-grey clay may either be regarded as the lower limit of a

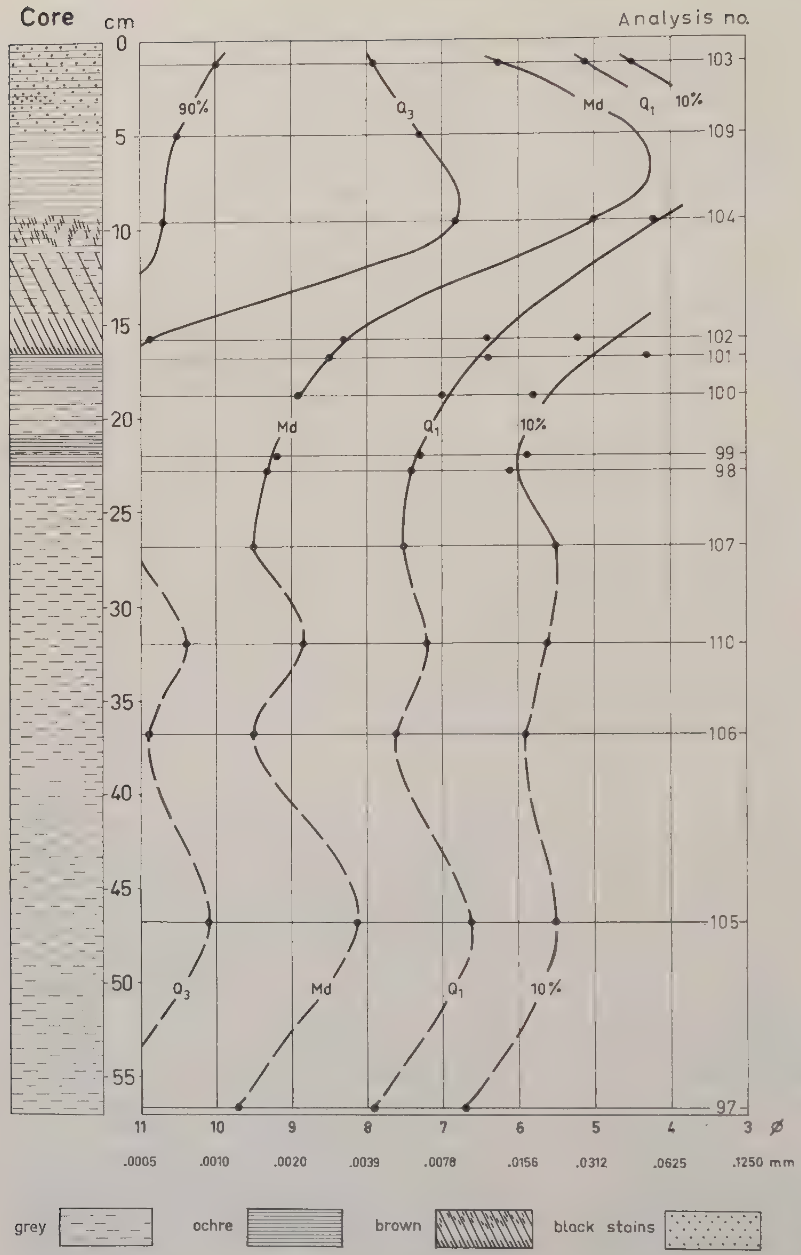


Fig. 135. Core from a depth of 26 m in section *C* east of Stora Röknen island. Grain-size distribution according to 13 analyses.

secondary chemical action on the total sediment, or as showing changes in the conditions during the sedimentation. The simultaneous change of grain size with the colour favours the latter alternative. The lower, grey part can be interpreted as a Glacial clay and the upper surface of this—the hard ochre-coloured stratum—was chemically affected before the upper part settled. This part must then be regarded as originating from suspended material washed out from Stora Röknen island, when it came above the water surface during the late Yoldia stage of the basin. When most of the fine material of the island had been washed

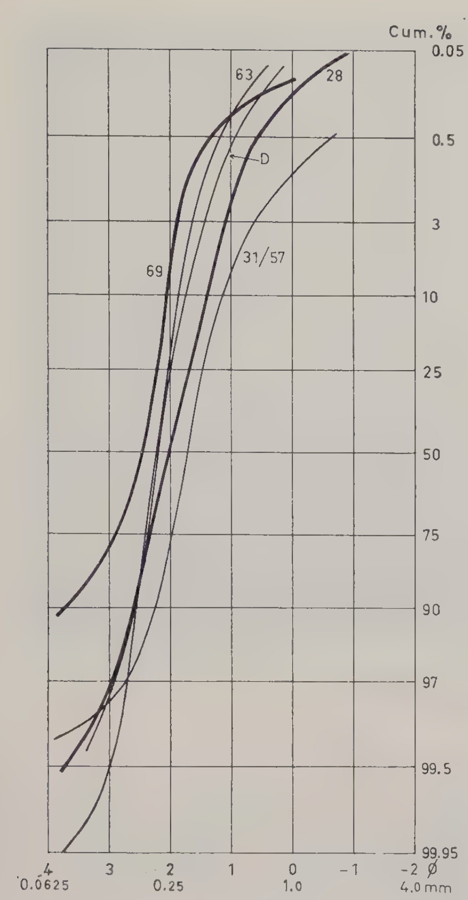


Fig. 136. Grain-size distributions of sand samples from the ridge east of Stora Röknen island (28, 69/61, heavier graphs) compared with samples from nearshore areas and a fluviatile sediment (*D*).

out, the rate of sedimentation rapidly decreased.

The spatial grain-size distribution of the superficial samples reported may seem somewhat confused, especially on account of the occurrence of gravel and scattered boulders deposited on the bottom. The particles of gravel size have been found mainly to consist of iron precipitates in deeper areas, and they can be regarded as having been formed *in situ*. (Not only gravel was found but also real crusts with a thickness of about 1 cm.) The real gravel was, as a rule, very angular, and can be presumed to have been transported in floes of lake ice. The scattered boulders may also have been transported by lake ice, and probably by ice jams over the low southern parts of the island during the submergence before the cutting of the outlet at Motala (cf. Fig. 7).

The fine sediments and the sand can be regarded as Glacial and Late-Glacial sediments covered by the most mobile bed-load fractions emanating from the shore.

The sediments of the central ridge

Sand cover and a deposit of coarse material

The first samples from the ridge were taken in 1961 with the Pettersson dredge. The dredge came up filled with a fresh well-sorted light-brown sand, without any traces of organic matter and iron precipitates. In a few cases there was some fine sediment in the lowest part of the dredge.

The distribution of two sand samples are shown in Fig. 136 (28/61 and 69/61). The first sample was taken just south of section D at a depth of 14.8 m, and the other one about 300 m further to the south along the western margin of the crest and at a depth of 15.5 m. The latter sample contained two gravel particles with a weight of 0.30 and 2.99 g. They are not included in the distribution of Fig. 136. The total weight of the dried sample was 444 g.

The first sample contains almost no particles finer than 0.06 mm (4 Ø) and about 1 % coarser than 0.6 mm (0.75 Ø). The other sample is somewhat finer; there is only 1 % coarser than 0.3 mm (1.75 Ø) and about 8 % finer than 0.06 mm. The distributions are compared with two samples of sediments from shallow areas connected with shores (Fig. 136), and with one sample of a fluvial sediment from the river Klarälven (Sundborg 1956, Fig. 48, sample D). Sample 31/57 is from the area off the Svedudden ness and was taken at a depth of 12 m, and sample 63/61 is the one from 0.9 m off the Stora Röknen shore. The five samples have an identical type of distribution; they are well-sorted and their main part lies within the interval 2 ± 0.5 Ø. In a fluvial sediment a distribution of this type indicates conditions at the transition between bed load and suspended load (Sundborg 1956, p. 279). In shore sediments this distribution can be expected in the outer part of the nearshore zone.

In 1962 a diving operation was carried out in a limited area on the crest at the south-

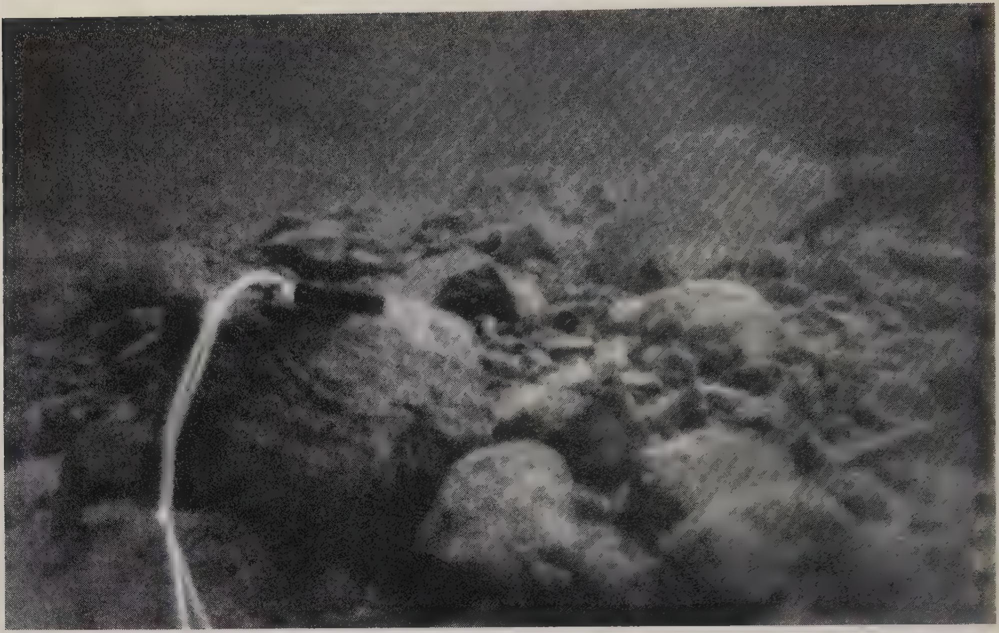


Fig. 137. Heap of moraine blocks and boulders on the ridge east of Stora Röknen island. Depth 17 m. A compass in the foreground. The block in the background has a size of about 1 m. Aug. 1962.

eastern end of the ridge. There were wide, rather flat areas of sand, the depth was about 17 m, and the bottom was rippled by oscillatory ripple marks (Fig. 138). The ripples had a crest-to-crest length of about 20 cm, and their height was about 3 cm. The ripple marks ran in an average direction of N.120°E, that is to say, perpendicular to the maximum fetch length.

A local area of boulders and blocks near the eastern margin of the ridge was also examined (Fig. 137). The existence of this area had been indicated by echo-soundings. This coarse material was angular and lay in a large low heap without any predominating orientation of the particles. There was no cover of sand or any other sediment; the stones were as clean as if they had been dropped only a short time ago. Of importance for the interpretation of the origin of this material is the fact that examination of the margin of the heap showed that it rests on the sand that covers the ridge. Therefore we can say that the material is a coarse moraine dropped from the bottom of an iceberg, and that the sand is an original glacifluvial deposit. Why the sand cannot be regarded as having been washed out of a moraine will be discussed further on. Here we shall only say that from the echo-soundings and the sediment sampling it is certain that the coarse morainic material is an isolated phenomenon but that more dropped moraine may exist.

Sediment cores

It was found difficult to obtain a sediment core from the ridge with the small piston core sampler. The sampler did not penetrate the sand deep enough during the falling moment and thereby fell over. However, one sample with a length of 58.5 cm was obtained in 1961 from the flat part of the crest in section D. The sand cover was only 4 cm, and below that there was a stratified fine sediment.

The lower part of this core (Fig. 139) up to the –27.5 cm level consists of a grey, varved, typical Glacial clay with thin, dark winter parts and thicker, light summer parts in the strata. The variation of the thickness of the varves from one year to another is small. The average thickness in the upper third of the varved part is somewhat less than in the lower parts (0.9, 1.3, and 1.1 cm). The grey varves

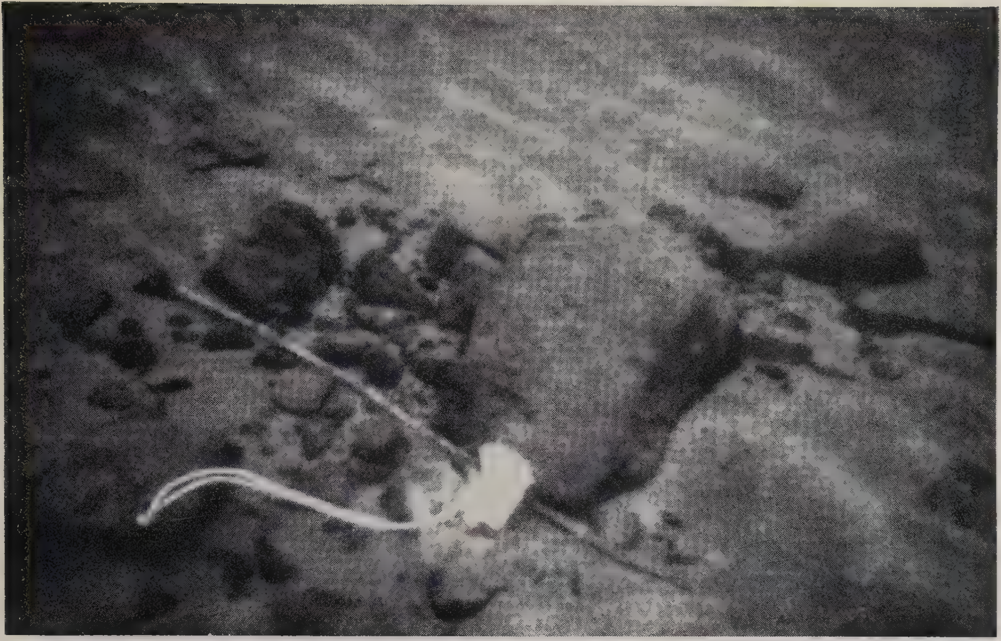


Fig. 138. Moraine boulders and cobbles on sand in the same area as that of Fig. 137. The stick in the foreground has a length of 1 m, and lies parallel to the ripples.

end abruptly at –27.5 cm, and they are followed by a brown, stratified sediment of rather thick, alternating light and dark strata, without visible intra-grading. It thereby has a banded appearance. The light strata were interpreted as summer deposits (S), and the dark ones as winter deposits (W), in conformity with varved clay (although the W-strata are often the thicker). In this series there are two thin strata (1 mm) with coarser particles. Both of them are found in the upper part of the dark strata.

The abrupt change from the varved clay to the brown sediment indicates a sudden change in the conditions of the sedimentation. With regard to the position of the locality in relation to the ice margin during the ice recession, the transition could be connected with the lowering of the Baltic Ice Lake to the sea or with the inflow of salt water from the sea after the lowering (cf. Bergsten 1943, Fig. 78, and G. Lundqvist 1961, the map). The banded appearance and the lack of intra-grading of the upper part resemble the sediments classified as symmict clay by Sauramo (1924), and regarded as having been deposited in brackish or salt water.

Seventeen grain-size analyses were made on strata of different levels. Summer and winter parts of varves were separated. The grain-size distribution of the core is illustrated by the characteristics of the median and the quartiles, as in the cores from the central section of the lake, and here the grain size of the 1, 5, 10, 90 and 95th percentages are also shown. There is a gap in the analyses between the –15 and –25 cm levels, because this part was taken out and given to Dr M.-B. Florin in an attempt to identify any micro-fossils. No identifiable fossils were found.

In the lower part up to the transition zone there is a clear difference in grain size between the light and the dark parts of the strata. The median of the latter coincides with the first quartile of the former within the same varve. In the transition zone there is no difference between dark and light parts, which seems to be due to an increase in grain size of the dark ones.

As regards the light strata, there is a continuous decrease to the –12.5 cm level. Within the grey varved clay the median decreases linearly (in the logarithmic scale) with a rate of 0.43 phi units/10 cm, or 0.038 phi units/

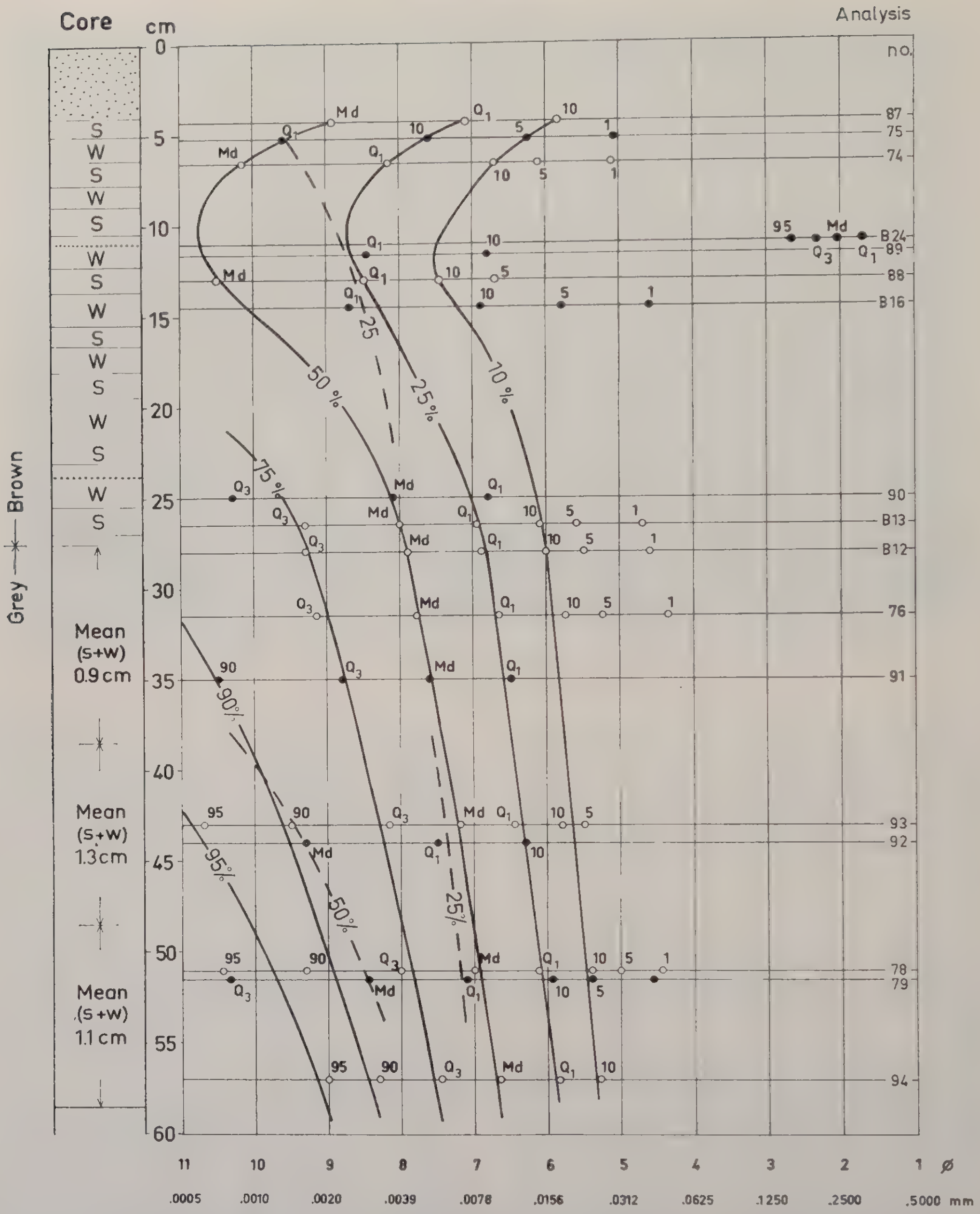


Fig. 139. Core from the ridge east of Stora Röknen island, taken at a depth of 20 m. Mean thickness of varves in the lower part, actual thickness of strata in the upper part. (S) light stratum, (W) dark statum, sand shown by dots. Grain-size distribution according to 17 analyses (1961). Open circles light strata, dots dark strata. Grain-size variation with evel of light strata shown by solid lines.

varve. The rate increases in the transition zone and in the brown sediment. In the uppermost centimetres there is again an increase of the grain size.

At the –57 cm level the grain-size distribution is quite symmetrical, but with higher levels the distributions become more skewed because of an increasing tail of fine particles.

There seems to be the same tendency for the grain size to decrease in the lower dark strata.

The distribution of the thin coarse layer at the –11 cm level was separated (B 24). The median lies at 2 Ø, and this sediment is thus comparable with the covering sand.

After discussion with Dr Landergren, of the Swedish Geological Survey, it was thought worth while to make a new attempt to get cores from the ridge for the determination of the boron and chlorine content, which might tell us whether the upper part of the stratified sediment was deposited in brackish (salt) water or fresh water.

In the summer of 1962 we were unlucky with the weather, and the core sampler was also lost at a depth of 38 m but was rescued again, thanks to good diving work by Mr Alexanderssson. On one of the last few days of the field season there was a calm night, and three cores were taken at the same anchorage between the sections E and D, and at a depth of 22 m. One of the cores with a length of 99 cm was immediately cut into slices with a thickness of 2 cm and put into tight plastic boxes exactly fitting the samples. It was noted that there was a change of colour from brown to grey at about 17 cm from the top of the core. The samples placed in the boxes were used for the determination of chlorine, boron, pore-water content, and grain-size analyses.

The following preparation of the samples was carried out in the Geomorphological Laboratory at the Institute of Geography in Uppsala. Half of the sample (about 30 g) was taken from the plastic box, placed on a plastic foil (0.6 mg) and weighed. Then the sediment was placed in a centrifugal tube and vibrated to a thick suspension, together with 10 ml distilled water, and was then centrifuged. Five millilitres of the water on top of the sediment was pipetted and put into a plastic tube, which was sealed. On these water samples the chlorine content of the pore water was determined.

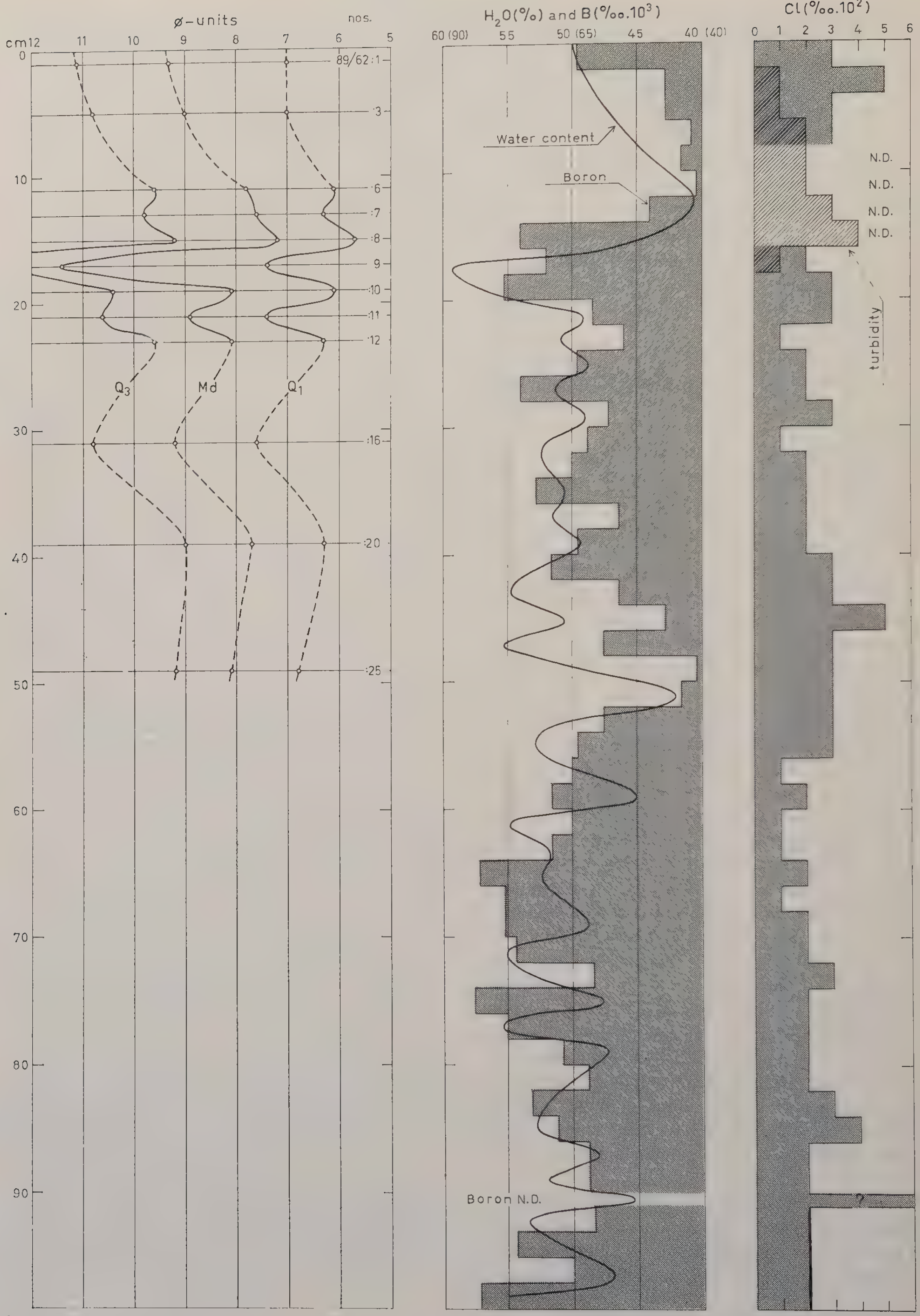
The clay left in the centrifugal tube was again suspended with distilled water, poured into a porcelain vessel, dried at 105°C, and then weighed. In this way the pore-water content was determined.

In eight samples all the matter did not settle, and there was still a colloid suspension. The turbidity of the pipetted suspension was subjectively estimated and graded in five classes, zero being clear water and 4 the most turbid. Even after four weeks there was still a suspension in the tubes.

For the determination of the boron content Dr Landergren wanted the fractions finer than 3 μ. Ten grams of each original sample were dispersed with 0.0034 g of $Na_2(COO)_2$ in 500 cm^3 of distilled water. The dispersion was made in acryl-plastic cylinders by turning them in a machine for 4 hours. Then the cylinders were placed in a constant temperature (20°C) for ten hours, and the upper 12 cm of the suspension was pipetted (about 328 cm^3). This suspension of particles finer than about 3 μ was centrifuged, the water decanted, and the sediment placed in plastic tubes.

The chemical analyses were carried out in the Geochemical Laboratory of the Swedish Geological Survey. The content of chlorine was determined by Dr F. Manheim. This content, expressed in ‰ of the original pore-water, was corrected in relation to a standard sample of 0.01 ‰. The boron content was spectrochemically determined by Dr Landergren and expressed in ‰ by weight of the dried sediment. The probable error of the method is $\leqq$ 10 % of the boron concentration (Landergren 1945*a*, p. 7).

The distributions of the characteristics determined are illustrated in Fig. 140. As the transition from the brown to the grey sediment was at about the – 17 cm level, the grain-size analyses were concentrated to this zone. The relationships that can be noted are all found in the upper part down to about the –18 cm level. The grain size increases down to –15 cm, and there is a sudden large drop at –17 cm. In the next sample it has increased again, and then the grain size seems to oscillate without any specific tendency. The pore-water content expressed as a percentage of the weight of the original wet sample continuously decreases from the top down to –13 cm (50 to 41 %), then suddenly increases down to –17 cm (59 %), and then from –21 cm oscillates around 50 %. The turbidity of the centrifuged samples increases down to the –15 cm level, then drops, and from –19 cm and downwards the water is clear. The boron content lies at 0.064 ‰ in the surface sample, then between 0.050 to 0.041 ‰ down to –13 cm, and then rises to over 0.070 ‰. From –19 cm to –50 cm there is a tendency to decrease. There are no specific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:9
:10
:11
:12
:16
:20
:25
Q_3
Md
Q_1
H_2O(%) and B(‰.10³)
60 (90)
55
50 (65)
45
40 (40)
Water content
Boron
Boron N.D.
Cl(‰.10²)
0
1
2
3
4
5
6
N.D.
N.D.
N.D.
N.D.
turbidity
?

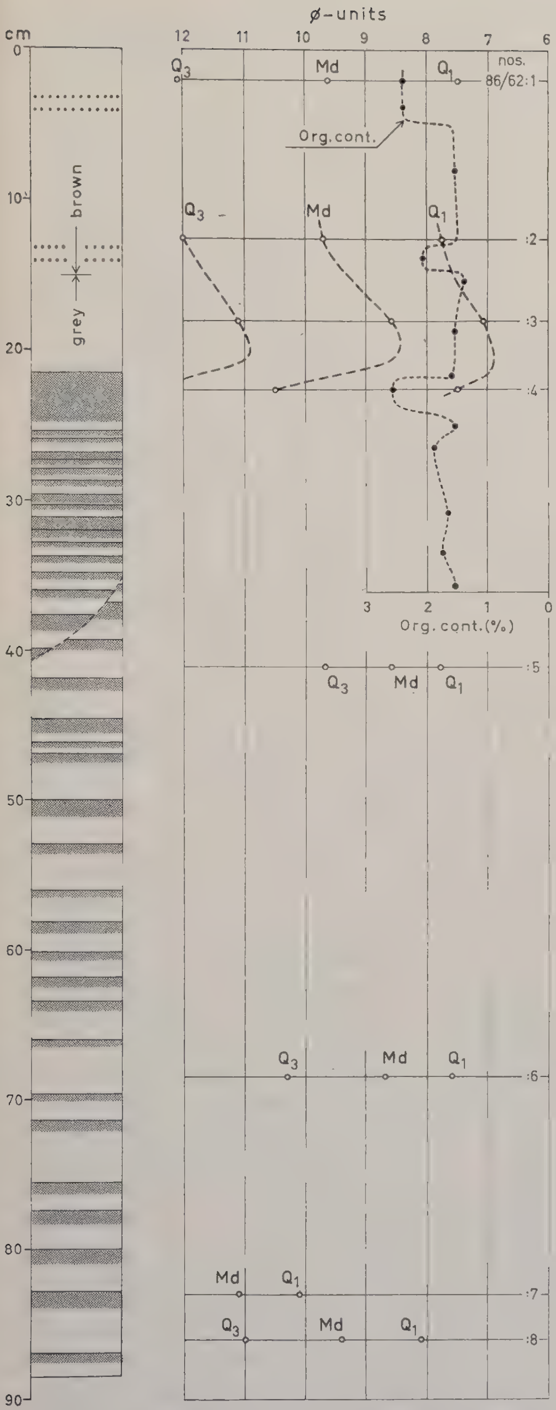


Fig. 141. Parallel core to the one of Fig. 140. Length of core 88.4 cm. Dark parts of varves shaded. Thin sandy strata marked by dots.

Fig. 140. Analyses of a core taken in 1962 on the southern part of the ridge east of Stora Röknen island. Depth 22 m. Length of core 99 cm. Scale of boron content in parenthesis. N.D. = not determined.

tendencies in the variation of the chlorine content, which is low.

According to Landergren (1945 *b*, Fig. 6), there is a linear relationship between the boron content of marine clays and the salinity of the water in which the sedimentation occurred. A boron content of 0.04 ‰ corresponds to a salinity of about 10 ‰, and a boron content of 0.08 ‰ to a salinity of about 20 ‰. The boron content of the Vättern sediment may not only be due to the salinity of the water during the sedimentation, but also to the marine origin of a part of the clay particles—they may originate from Paleozoic sediments abraded by the ice. With regard to the small variation of the boron content in the analyses of the core, no change of salinity is indicated. The somewhat lower content in the upper part may be due to a variation in the sources of the sediment.

As regards the grain-size distribution, it must be noted that the analyses were not made on specific strata but on slices with a uniform thickness.

The stratigraphy had to be determined from one of the other cores from the same anchorage, which was not cut into slices when sampled. This core, which had a length of 88 cm, is shown in Fig. 141. Up to the –21.5 cm level there is a varved clay. It ends with a thick dark stratum. Above that the sediment has no visible stratification, except four thin, sandy strata. The colour changes from grey to brown at –15 cm. Two grain-size analyses of a varve at the bottom of the core show the regular difference between the light summer part of the varve and the dark winter part. The analyses from the –68, –41 and –18 cm levels have almost the same median values. The sediment of the thick, dark stratum is considerably finer. Analyses of the organic content, expressed by percentual loss of weight of dried samples after heating to 500°C for 4 hours, were made on the upper 35 cm. The analyses of the varved clay were made on the light parts of the varves up to the thick, dark stratum. The content lies at about 1.5 to 2.0 %, except for the middle part of this stratum, where it is 2.56 %, and near the top, where it is 2.4 %.

This core and the previous one can be correlated by the thick, dark stratum, which is distinguished by a finer grain size, a higher content of organic matter and a larger pore-water volume. It may be noted that in these cores, taken only about 500 m south of the core illustrated in Fig. 139, there is no upper part with the character of a symmict "banded" clay.

The mud of the deep areas

The mud or gyttja of the troughs and channels in the Stora Röknen area were preliminarily surveyed. Samples were taken with the dredge down to depths of 90 m, and with the core sampler down to depths of 60 m. The mud was found in all the deep areas below the slopes, but not on the slopes. The samplers were always filled with a black gelatinous mud, and thus its thickness can be estimated to be at least 1 m.

Preliminary analyses of the organic content have shown values between 4 to 15 % by weight of the dried sediment. The inorganic particles are finer than 4 Ø, and the median of the grain-size distributions lies around 8 Ø. That is to say, the distributions are similar to those of the Glacial clay. For further investigations on this mud a larger piston core sampler is needed, and the organic content should be analysed by a specialist on Quaternary fossils.

Conclusion

The problems as regards the sediments of the ridge east of Stora Röknen island cannot be entirely solved by the investigations carried out in 1960–62. It was thought that further field studies would not be likely to add more information of great importance with the equipment available. A heavier core sampler, by which the varved clay can be penetrated, is needed.

From the present knowledge of the sediments the following interpretation of the late stages of the Glacial conditions may be proposed. When the ice margin was situated at about the latitude of the ridge, large, marginal, glacifluvial deposits were formed in the Vättern basin east and west of the lake, and they were partly formed in narrow ice estuaries (cf. Bergsten 1943, p. 181). It seems possible that the ridge also marks an estuary of this type, in which glacifluvial sediments were deposited. The varved clay indicates conditions of low velocities, which, judging from the occurrence of the covering sand, must suddenly have changed.

The covering sand cannot have been formed by erosion or abrasion of the fine sediments. The known change in the basin that can be connected with a radical change in the drainage conditions is the lowering of the Baltic Ice Lake to the sea. Not only the lowering, as such, but also the tremendous flow south of the ice margin towards the Karlsborg area must have influenced the drainage pattern. These changes may have caused an intensified intraglacial and superglacial drainage towards the estuary from the surrounding higher areas, and these flows would have brought the sand.

Atfer the lowering to the sea the ice recession was rapid (cf. Lundqvist 1961, the map). With no ice in the deep channels, the dense suspended loads of glacifluvial flows could follow these channels, and no fine sediments settled on the crest of the ridge.

Submerged shores of the central and southern parts of the lake

The Lowest Shoreline

North of Visingsö Island

The Flisen area is the most northerly locality at which morphological traces of the lowest shoreline were searched for by echo-sounding and diving operations. The bottom of this area is very even, and no discontinuities indicating shorelines were found by echo-sounding. Wave-washed sediments with gravel and cobbles on clayey moraine were examined down to depths of 11 m during the diving operations, but these surfaces were gradually replaced by re-deposited fine sediments. It is certain that the lowest shoreline of this area has at least been as low as 9–10 m below the present water level, but the lowest level cannot be stated.

North of Visingsö island there are large areas of wave-washed clayey moraine with a cover of cobbles and gravel. Because of the experience from the Flisen area, the lowest limit of wave wash was searched for along the eastern margin of the plateau, where the slope is rather steep towards the deep channel. The search was also carried out as far north as possible, because the particles washed out of the moraine were esti-

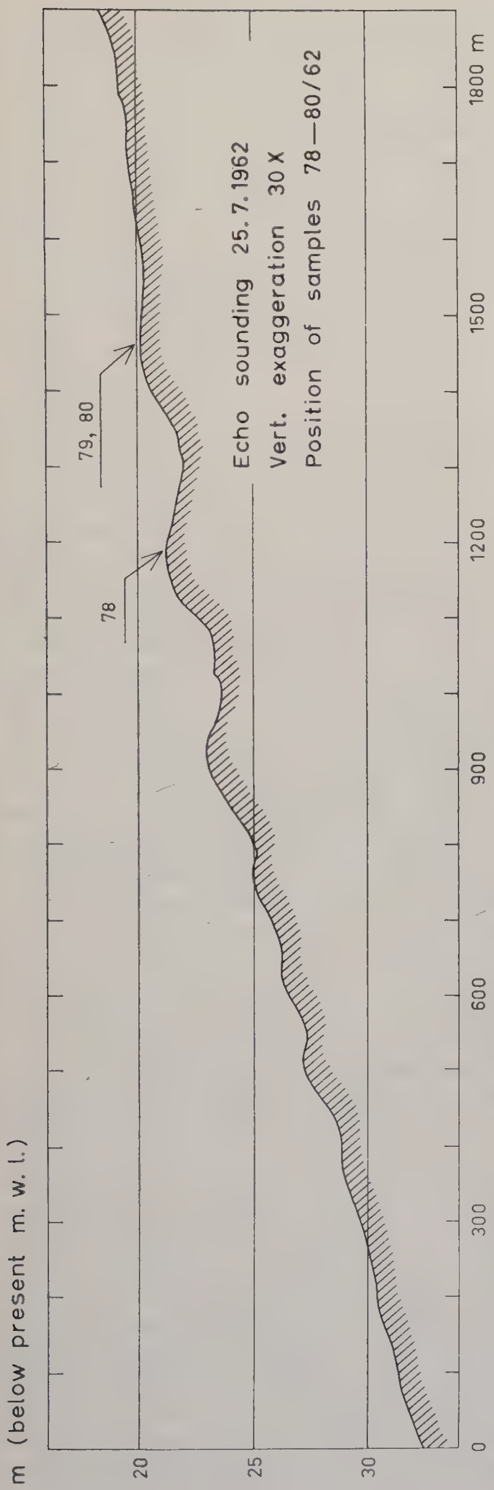


Fig. 142. Profile of submerged bar forms off Gränna harbour.

mated to have been transported southwards towards the large plateau of sand off the Sandudden ness. A sharp limit between wave-washed moraine and clayey moraine with a cover of clay was also found along this margin at the latitude N.58° 10′ during a diving operation in June 1962.

The examination of the bottom at this site started at a depth of 17 m. The bottom was covered with pebbles and gravel, and there was a surface crust of iron precipitates between the pebbles. In a section towards the east the depth increased from 17 m to 21 m over a distance of 35 m. At the depth of 21 m there was a sharp limit towards a clay surface. Only a few metres from this limit the thickness of the clay, which was rather soft, was more than 0.4 m. Nearest to the wave-washed surface there were some pebbles on top of the clay. No coarse particles were found in the clay. Because of the sharp limit, the fine sediment must be regarded as being of Glacial origin, and the 21 m level is the lowest level of intense wave wash.

The distribution of the scattered pebbles on the sediment surface makes it probable that the water level has never been lower than 21 m, but the lowest level may have been somewhat higher. According to the lowest shoreline predicted by extrapolation (E. Nilsson 1953, Pl. IV), the lowest level at this latitude should lie at about 23 m below the present water level. With respect to the possible precision of the extrapolation, this value agrees rather well with the observation of the lowest wave wash.

The Gränna Area

In this area a morphology interpreted as submerged subaquatic bars was noted during echo-soundings. The position of these bars is marked in Fig. 4, and an original echogram is reproduced in Fig. 43. The contour of the same profile has been re-drawn in Fig. 142.

The bottom was examined by diving, and at the same time samples were taken on the upper bars. There was a very thin cover of fine particles on the bottom and sand below that. The samples consist of well-sorted sand in the range 1–3 Ø and a few gravel particles (Table 26, nos. 78–80/62), and thus it is certainly of a nearshore origin. The samples are marked at the 20 and 21 m levels in Fig. 57.

The lowest shoreline at this latitude predicted by E. Nilsson lies at about 30 m below the present

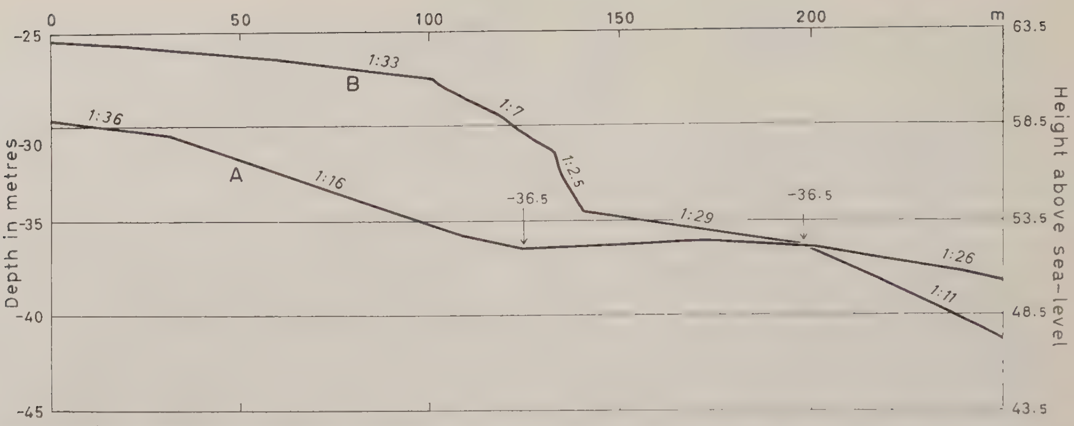


Fig. 143. Echo-sounded bottom profiles from the western slope of the abrasion platform off the Rosenlund cliffs

water level, that is, at the lower end of the series of bars in Fig. 142. The echo-sounded profile and the samples do not prove the lowest water level. The bars may, of course, have been formed during the submergence. There is also another problem concerning this morphology, viz. the dimensions of the bars above the 25 m level. These are much larger than any subaquatic bars of the present shores.

The same problem was faced by van Straaten (1959) in an investigation on echo-sounded "drowned bars" off the Rhône delta. Mainly because of their dimensions—average length 400 to 1000 m—it was possible that these bars were submerged barriers. However, bars can be presumed to have a greater chance of being preserved during submergence than barriers which are passed by the shoreline, and van Straaten finds it more probable that the forms were originally normal submarine bars, which were widened during the submergence (op. cit., p. 259). In the Gränna area there are thus three possible alternatives; (*a*) the lowest shoreline lies below the bars at about 30 m, and the bars were formed during the submergence, (*b*) the lowest shoreline lies at about 25 m, and the upper large bars were formed during the submergence, or they are submerged barriers, (*c*) all the ridges are subaquatic bars, and the water level was never lower than about 20 m below the present water level. It should also be noted that the morphology may in part have been formed during the stage of emergence preceding the submergence.

The last alternative (*c*) is not very likely because of the lowest level of wave wash observed north of Visingsö island. Thus it seems most probable that the lowest shore level lies at 25 m below the present water level or lower.

The southern end of the lake and the relationship between observed shore features

No morphological traces of a shoreline as low as the one predicted by E. Nilsson (43 m below the

Table 24. Calculations on the lowest shoreline of the Vättern basin (heights in metres above sea-level). The relation to the present lake level is shown in parenthesis.

Locality	Distance[1] (km)	Calculated shore levels				Observations
Jönköping	0	*46 (—42.5)*	*46 (—42.5)*	*52 (—36.5)*	*52 (—36.5)*	52 (—36.5) echo-soundings
Gränna	29	58.7 (—29.8)	59.1 (—29.4)	63.4 (—25.1)	63.8 (—24.7)	63.5 (—25.0) bars/barriers?
N.58°10′	43	64.8 (—23.7)	65.4 (—23.1)	68.9 (—19.6)	69.5 (—19.0)	67.5 (—21.0) lowest wave wash
Tibon	135	*105 (+16.5)*	*107 (+18.5)*	*105 (+16.5)*	*107 (+18.5)*	
Shoreline gradient (m/km)		0.437	0.452	0.393	0.407	

[1] According to E. Nilsson's shoreline diagram (1953, Pl. IV).

present water level) were found by echo-soundings of the bottom slopes at the southern end of the lake. In fact, there was only one indication of a very low shoreline, and that was along the western flank of the abrasion platform off the Rosenlund cliffs (cf. Fig. 74). The general absence of traces of shorelines can be ascribed to the sedimentation, which has smoothed the profiles.

On the slope of the abrasion platform, which is covered with sand, there are two echo-sounded parallel profiles directed towards the WNW. and situated 400 m apart. In the lower part of these profiles there is only one hack, and in both of them this lies at the same level (Fig. 143). In the outer profile (*B*), which runs about 500 m SSE. of the outer end of the abrasion platform, there is a well-marked scarp. This profile may be compared with the one of the sandstone cliff on Visingsö island, which has been drawn with the same vertical exaggeration (Fig. 100). The other profile is not so accentuated, but there is a distinct bench with a width of 75 m.

The profiles are situated in an area where a shoreline morphology could be expected to be preserved, because this flank was exposed to large waves and beach-drifting, and the transport of the sediments abraded on the slope has been towards the Jönköping bay. Thus little sedimentation by suspended load can be expected, and the profile has not been smoothed.

If it is accepted that these profiles show the lowest shoreline, the lowest water level can be estimated as 36–37 m below the present water level, which is 6 m higher than E. Nilsson's lowest level.

Is the level indicated by the presumed shoreline at the southern end of the lake combinable with the observations in the Gränna area and north of Visingsö island and with the threshold of this lake stage? During the lowest stage the lake was isolated from the sea by a threshold at Lake Tibon NNE. of Askersund, which is situated at 105 m above sea-level. However, it seems uncertain whether this level or a 2 m higher water level should be used (cf. E. Nilsson 1953, Fig. 25 and Pl. IV), but this difference of 2 m is of minor importance as regards the southern part of the lake.

In Table 24 the water levels of the localities of special interest are shown. They have been calculated on the assumption of a lowest level at the southern end of the lake of 46 m and of 52 m above sea-level and of the two alternative levels of 105 m and 107 at the Tibon threshold.

With the assumption of a 52 m level at Jönköping, the wave-wash limit observed north of Visingsö island lies 1.5 to 2.0 m below the lowest shoreline which can be regarded as possible with respect to the activity of breakers. The same shoreline intersects the profile of the Gränna area at the depth of 25 m, that is to say, between the bars of normal dimensions and the large bars (barriers?). As I mentioned before, the assumption of a 46 m level at Jönköping involves a shoreline which intersects the Gränna profile at a depth of 30 m and which runs 1 to 1.5 m below the limit of wave wash north of Visingsö island.

With regard to the fact that E. Nilsson's shore level at the southern end of the lake is the result of an extrapolation over 100 km within the basin, the difference between the two shorelines calculated is small, and it seems most likely that all the submerged features described belong to the lowest shoreline of the basin. The shoreline gradients arrived at by assuming the 52 m level in Jönköping will, according to formula (1), change the age of the lowest shoreline from about 7400 B.C. to 7300–7100 B.C. (cf. Norrman 1964, Fig. 6).

Higher submerged shores of the southern part of the lake

The echo-soundings have shown occasional steps in the bottom slope profiles, but not until a rather high level is there a break in the profiles. It can be found in most of the echograms. This type of break is well illustrated by two profiles from the western shore, north of the Jönköping bay (Figs. 144 and 145). The slope of the deep channel is rather even, until there is abruptly the margin of a platform at a depth of about 10 m. In this area along the western shore the platform may reflect a bedrock terrace of the Visingsö Series, although the bedrock is not exposed. However, a margin is found at about the same level in several other profiles around the southern end of the lake. It is, for example, illustrated by the profile from the Huskvarna bay in Fig. 113, and in the map of this bay (Fig. 147). In Table 25 the level of the break in the profiles is summarized.

Shoreward of the break there are wide platforms and gentle slopes. The broadest platform is that off the Rosenlund cliffs. In its central part

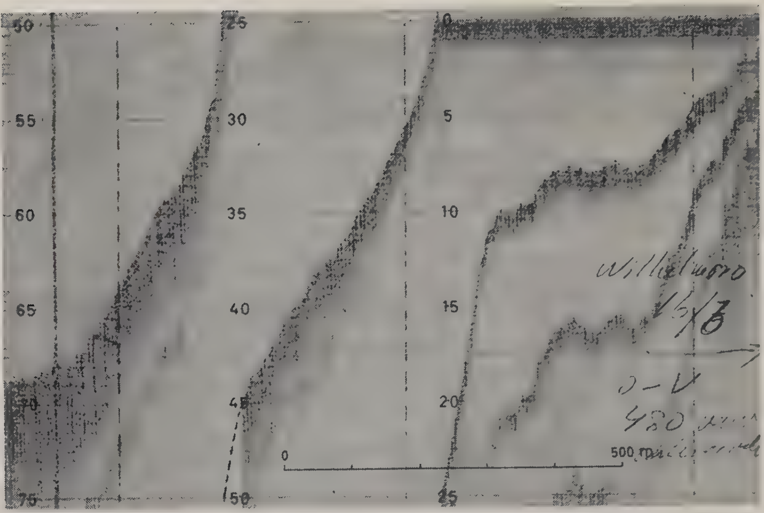


Fig. 144. Echogram of the bottom profile in the Vilhelmsro area. Direction of section E.—W.

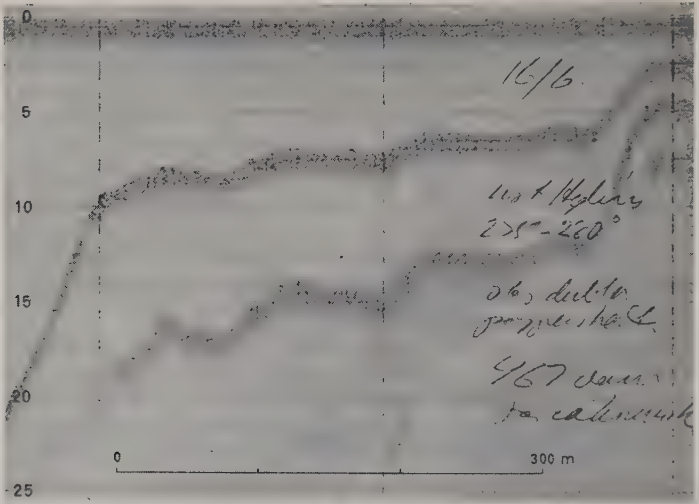


Fig. 145. Echogram of the bottom profile at Kaptensbo. Direction N. 95° E.

the bottom over wide areas lies at a depth of about 7 m (Fig. 146). The same depth characterizes the platforms of the western shore.

The 10 m level was passed by the shore during two different periods, which may both have contributed to its specific morphology. The first period was about 8200 B.C., when the Baltic Ice Lake had been lowered to the stage of the Yoldia sea (cf. Fig. 7). According to E. Nilsson (1953, Pl. IV), this Yoldia shore was at the southern end of the lake at 77 to 79 m above sea-level, that is to say, at a present depth of 8.5 to 10.5 m. Because of a rapid land uplift, this shore level was probably of a short duration, but the shoreline

Table 25. Characteristic depths (in metres) of platforms and margins at the southern end of Lake Vättern.

Locality	Margin	Platform
Vilhelmsro, western shore . . .	10	7.5—8
Kaptensbo, western shore . . .	9.5	6—8
Abrasion platfo m off the Rosenlund cliffs;		7—9
western flank, ...	10	
outer end, ...	11.5	
eastern flank	10	
Huskvarna bay;		Slope
western end, ...	10	
central part, ...	8.5	
eastern end	8	

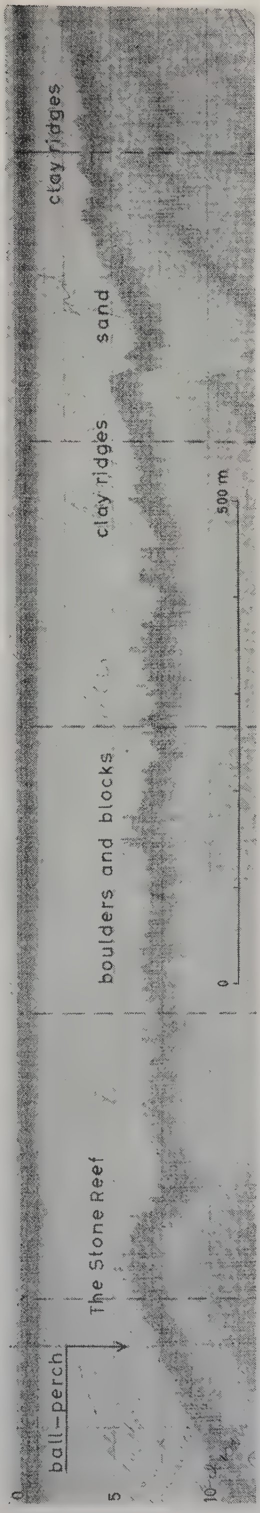


Fig. 146. Echogram of the bottom morphology of the abrasion platform off the Rosenlund cliffs. Section from the northern end of the platform towards the large ravine in the eastern part of the cliffs. The profile ends 105 m from the shoreline.

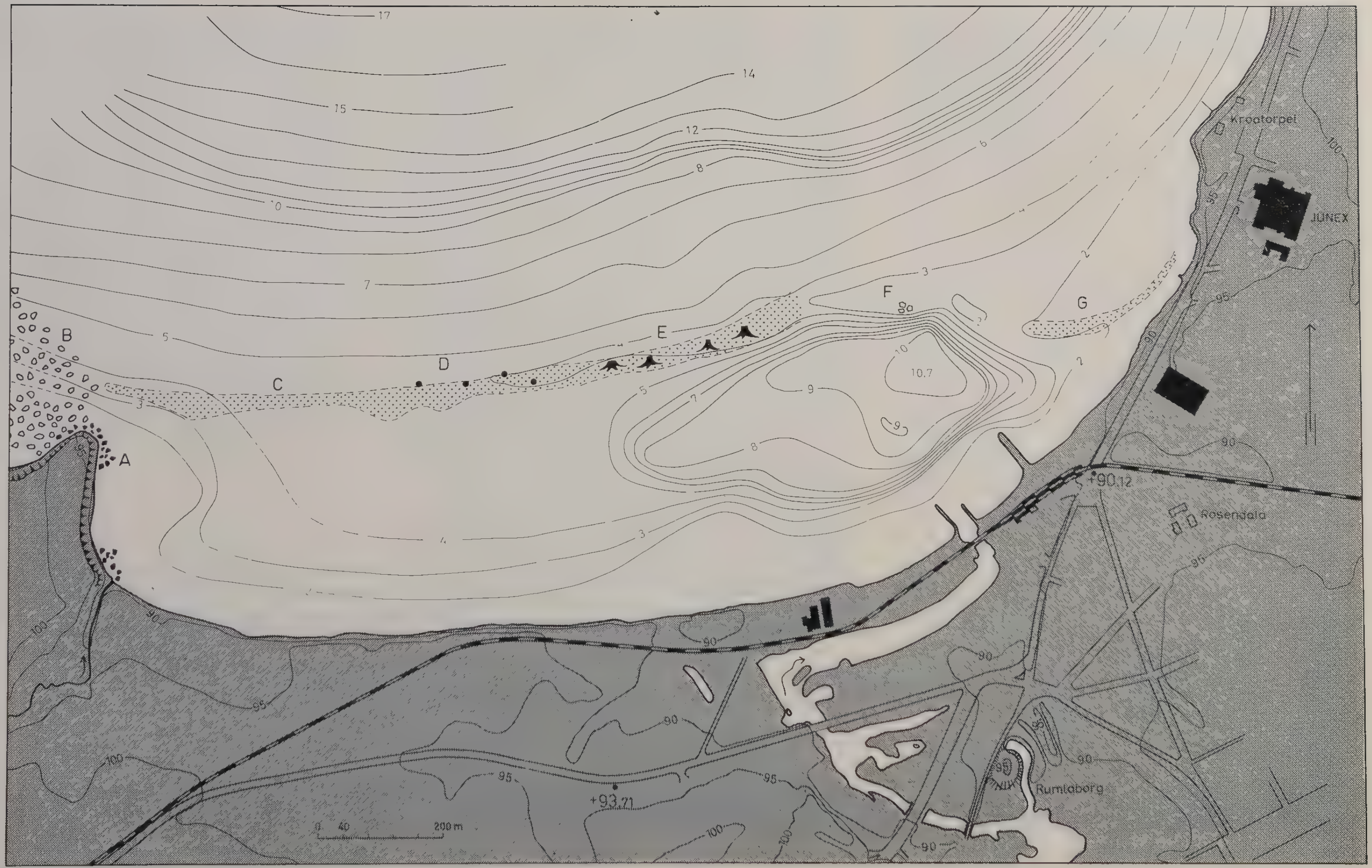
of the early Yoldia stage is well marked in the central and northern parts of the basin (cf. Bergsten 1943, E. Nilsson 1953). At the latitude of Gränna the shoreline is found at 3.5 m above the present water level, and in the central part of Omberg at about 22 m (E. Nilsson 1953, p. 202). It thus seems most probable that a shore platform was cut at the exposed southern end of the lake.

When exactly the 10 m level was again passed by the shoreline during the stage of submergence is uncertain. Palynologically dated levels of lagoon deposits at the southern end of the lake form a consistent smooth graph back to 2000 B.C., when, according to these datings, the lake level at the southern end was about 5 m lower than the present level (Granlund 1934 *a*). If this graph is extrapolated back in time, the 10 m level is reached at about 4000 B.C. However, the oldest palynological levels, dated 2500–2700 B.C., lie in the depth interval of 9–11 m. The absolute dating of these levels may be erroneous, but they still indicate that before 2000 B.C. there was a period of slow submergence, preceded by a period of "normal" or even rapid submergence and followed by a period of rapid submergence. These variations in the rate of submergence were most probably due to climatic effects on the water level (cf. Norrman 1964) and not to variations in the rate of the uplift.

Through the lowering of the water stage during a general trend of submergence the abrasion will act in a limited height interval for an unusually long time, and an abrasion platform can be expected to be formed. As regards the platform level at the southern end of Lake Vättern, an explanation involving variation of water stage or any other explanation must be hypothetical with our present knowledge of the age of the lagoon sediments. However, inshore of the barrier zone of the Huskvarna bay there are organic sediments down to a depth of at least 20 m, and it is hoped that these sediments will soon be dated.

There is still much to be done as regards the chronology and morphology of the submerged areas at the southern end of the lake, and the Huskvarna bay can be judged to be the most interesting of the areas. Some features of the morphology of this bay will finally be discussed.

A bathymetric map of the bay was published by Granlund (1934 *a*), but this map is partly erroneous. The bay was investigated by echo-soundings and diving operations in 1961 and 1962, and the map of Fig. 147 is mainly based on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A
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F
G
Kroatorpet
JUNEX
Rosendala
Rumlaborg
+90.12
+93.71
90
95
100
0 40 200 m

Fig. 147. Map of the morphology of the Huskvarna bay according to echo-soundings and diving operations in the period 1961–63. Areas of superficial organic deposits of the bay dotted. *A*, boulders and blocks projecting above the water surface. *B*, the Stone Reef—boulders and blocks. *C*, area with peat and logs. *D*, area with peat and artefacts (black spots) investigated by Alexandersson in 1963. *E*, area with pine stumps. *F*, modern stone caisson. *G*, peat.

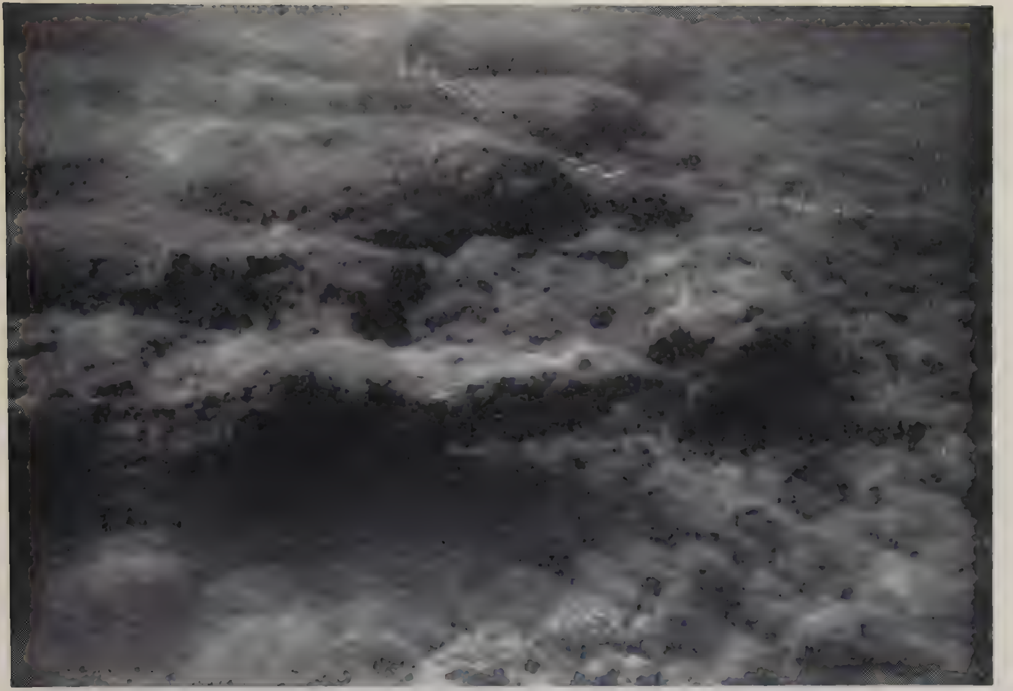

Fig. 148. The front of the peat ridge in area *C* (cf. Fig. 147). Note the furrows in the peat surface and the end of a log in the foreground. The bottom in front of the ridge consists of cobbles covered with algae.

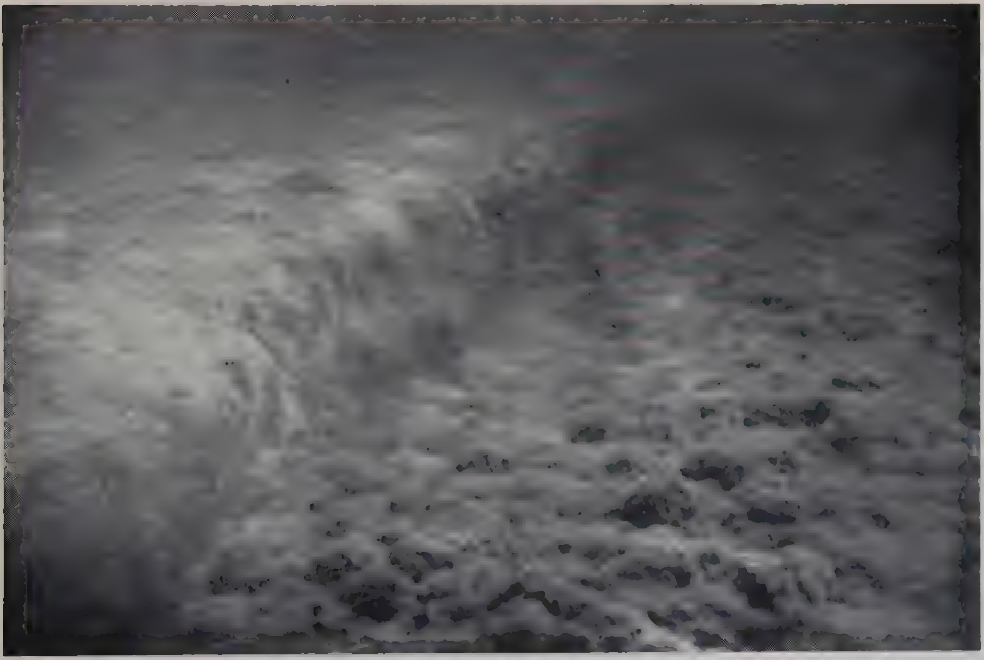

Fig. 149. The back of the peat ridge in area *C*. Its height is about 0.5 m. Cobbles covered with algae in the foreground.

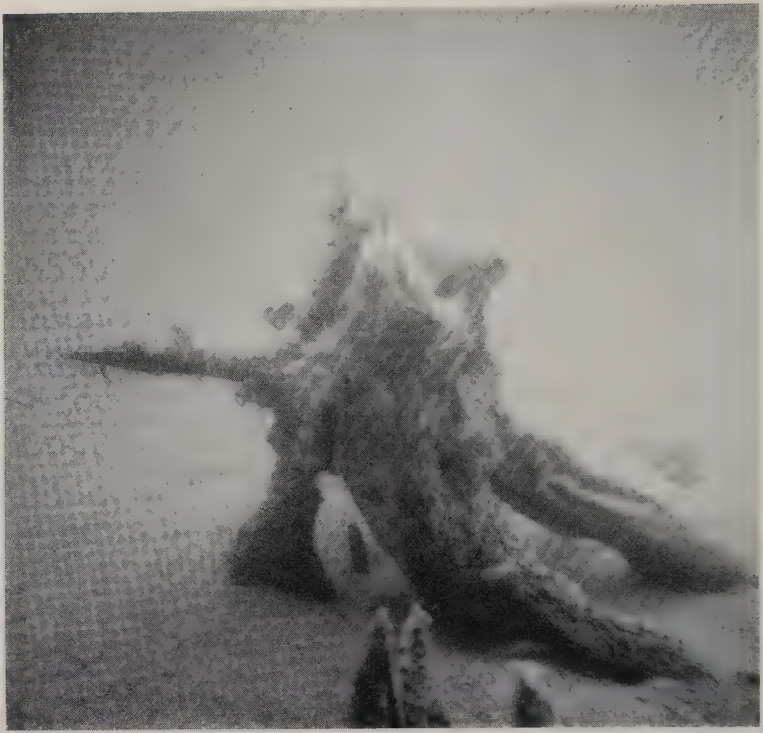

Fig. 150. A stump of Scotch pine on the bottom of the Huskvarna bay (area *E* in Fig. 147). July 1961.

these investigations. The most striking feature of the bottom morphology is the enclosed depression off the outlet of the Huskvarna river. A profile of the depression is shown in Fig. 113. This hollow must be regarded as having formed a part of the river valley during the low stage of the lake. It may be noted that the bottom has a depth of 9–10 m over a large area. The bottom sediments are muddy. The hollow has not been filled up by the river, because there is almost no bedload transport. Towards the north-west it is bounded by a sand slope of distal-slope character.

Another feature of great interest is a band of organic deposits and submerged tree stumps, which fringes the bay at the 4 m level. The first note on these deposits was published by Weiler (1926), who mentioned a site containing oak stumps corresponding to the area *E* of Fig. 147. From the same area peat was sampled by Granlund and dated by pollen analyses to about 1500–1000 B.C. (1934 *a*, p. 23). The distribution of the deposits was not mapped.

Through the diving operations in 1961 and 1962 it was found that there was a continuous band of somewhat variable deposits all along the central and western part of the bay. In the western part (*C*) there is a well-defined flat ridge of peat with a height of about 0.5 m. There are some logs in the peat lying perpendicular to the direction of the ridge (Figs. 148 and 149), but no stumps. One log with a length of 8 m, taken up in 1962, was found to be half an oak log, split along its length. A part of this log was microscopically examined by Mr. T. Ingmar. In this western area there are distinct furrows in the peat surface, which run perpendicular to the ridge (cf. Fig. 148). These furrows may be a corrasion phenomenon.

In the central part (*D*) the continuity of the peat was uncertain because of a sand cover. In this area tree stumps were observed and also fallen pine trees with a length of 4 m and their roots in the peat. Along this area there were also low hummocks of boulders and cobbles. The area was further investigated in 1963 by my co-diver, Mr T. Alexandersson (1963), who found several traces of human activity in this area. The sites of his finds are marked by dots in Fig. 147. Some thin logs were found to be poles that had been driven into the bottom both in the peat ridge and in the surrounding gravel and sand. There were also remnants of open, timbered boxes, 2 m in square, filled with boulders. Eight boxes situated in two rows fairly parallel to the ridge seem to have been identified. These boxes can be estimated to have formed stone caissons which either belonged to a pier or a construction for fishing. The age of the finds is not discussed by Alexandersson. Because of the timbered construction, they are certainly more recent than the peat and the (tree) stumps.

In the eastern part (*E*) little peat is exposed, but there are many pine stumps (Fig. 150). Some of these stumps stand on low hummocks of boulders and cobbles. The age of one of the stumps of Scotch pine has been determined at the Carbon-14-Dating Laboratory in Stockholm, and its age was found to be 2840 ± 90 B.P. (analysis St-1056), that is to say, about 900 B.C., an age which agrees well with the rate of submergence, according to Granlund's datings.

As regards the species of the stumps, it may be mentioned that only Scotch pine and sallow were found, and no oak. Samples of the wood have been examined by Mr. T. Ingmar.

Just off the eastern shore of the bay there is another deposit of peat at a depth of 1 to 2 m. Its outer part was indicated by the echo-soundings. In connection with a preliminary survey for a new road bank this peat was examined by Mr. L.-K. Königsson and myself in the late autumn of 1962. It occupies a shallow depression, and its upper surface lies on a level with the surrounding sand and gravel bottom.

The peat of the bay is of the type formed in shore lagoons with a sedge flora (cf. Granlund 1934 *a*, Fig. 12). It rests on gravel and sand, and in front of the peat ridge there is an abrasion pavement of cobbles. It seems most probable that the peat was formed in a lagoon behind a beach ridge or barrier along the bay. No morphological indications of the outlet of the river Huskvarnaån were found by the echo-soundings or during the divings. It may have been situated between the two areas of peat.

During the submergence the barriers along the bay was abraded, but the peat behind the barrier was partly preserved. We do not know when this happened. The water level was probably well above the upper part of the stumps, because it seems improbable that any stumps would have been left on this exposed shore, if the abrasion had taken place when the water surface was on a level with the present peat ridge and the base of the stumps. Thus it can be estimated that the barrier was abraded after the Birth of Christ, when the water level was about 3 m lower than the present level, and before the barriers at Rumlaborg started to develop, which according to Granlund's datings occurred at about 1100 A.D.

Conclusion

Above a depth of 10 m there are several indications of abrasion platforms at the southern end of the lake, but below that level there is only one indication of a lowest shore level at a depth of 36–37 m. From the finds of a submerged peat ridge and pine stumps at a depth of 4 m along the outer part of the Huskvarna bay it seems probable that this bay was enclosed by a barrier in front of the peat ridge. Because of the age of one of the pine stumps, it can be estimated that the barrier was probably not abraded until after the Birth of Christ.

RESULTS

This paper was written with the aim of describing and analysing the shore and bottom morphology of a specific area, which means that it includes descriptive parts which cannot be summarized in terms of briefly specified results. However, it also includes investigations on specific problems, some of which yielded results of general applicability, while others were of importance for understanding the local developments. Among these results I want to draw special attention to the following.

As regards the methods, the investigations prove that diving operations make possible a rapid survey of the bottom morphology of nearshore areas, and that direct observation of the bottom is a necessary complement to echo-soundings. By the use of diving technique it is also possible to map morphological details with a precision which cannot be obtained only by sounding methods.

The investigations on the shores demonstrate a relationship between the morphology and the kinds of deposits and bedrocks, which, with regard to the morphological effects of the shore processes, can be referred to the four categories of frictional and cohesive soils, and sedimentary and crystalline rocks. The marked difference between the morphology of areas of submergence and emergence is proved by the predominant cliff morphology and the barriers of the former, and the wave-washed but generally little transformed slopes and the dunes of the latter.

The shoreline configuration formed by the shore processes can be classified in a form series of indentations and bays, and the orientations of these forms are proved to be mainly due to the fetch distribution. By map comparisons and new surveys it has been possible to estimate the recent shore retreat in the areas of most active abrasion in the southern parts of the lake.

The general retreating charachter of the shores also stamps the morphology of the beaches. The shore retreat is not only a result of submergence, but is also due to the destructive effect of the steep waves of the lake. This is proved by the morphology of the foreshore and the nearshore bottom. Characteristic features of the beaches are their small width, the absence of medium sand and finer fractions, and the marked step consisting of gravel and pebbles. The main part of the nearshore bottom of the lake consists of an abrasion pavement, but in some areas of wave-energy deficit there are also platforms of sand on a level below the normal breaking depth. The nearshore bottom morphology of bays with sand accretion by longshore drift is characterized by subaquatic bars, which can be regarded as the result of the destructive wave action.

It has been proved that corrasion is a most effective process for the abrasion of cohesive sediments. The resulting morphology much resembles that formed in fluviatile conditions. As regards the abrasion of the sedimentary rocks, the frost action seems to be of great importance.

The distribution of recently deposited nearshore sediments can be qualitatively related to the estimated wave-size distributions, and it has also been possible to determine the transition from bed-load sediments to suspended-load sediments.

The sediment distribution in the offshore areas involves the problems of former low water levels and glacifluvial sedimentation. It has been proved that the lowest shoreline of the basin, predicted by E. Nilsson (1953), approximately fits the distribution of the superficial sediments. Morphological traces of the lowest shore and higher submerged shorelines have been identified by echo-soundings, and have also been directly examined by diving operations.

Core samples from the central and northern parts of the lake prove that the bottom of the lake is covered with glacifluvial clay and that, except for the submerged areas, which have been passed by the shore zone, and the deep channels, which contain organic mud, the sedimentation since the ice recession has been insignificant.

Table 00. Characteristics of grain-size distributions referred to in text and figures. Numbers *(Nos.)* according to the register cards of the Geomorphological Laboratory. *Types:* (F) foreshore and cliff samples, (B) bottom samples taken by dredge and by hand, (C) bottom cores. *Depths* in metres. Modality *(Mod.):* (U) unimodal, (B) bimodal. Ø$_{16}$: phi value of 16th percentile (coarser than). % < 3Ø: percentage finer than 3 phi. M$_ø$: mean. Md$_ø$: median. $\sigma_ø$: dispersion. $\alpha_ø$: skewness.

Nos.	Type	Depth	Mod.	Ø$_{16}$	Ø$_{25}$	Ø$_{75}$	Ø$_{84}$	% < 3Ø	M$_ø$	Md$_ø$	$\sigma_ø$	$\alpha_ø$	Notes
1957													
1	F	—	B	—3.80	—0.70	0.75	0.95	0	—1,43	0.30	2.38	—0.73	Beach sediments, Visingsö island
2	F	—	U	0.60	0.75	1.30	1.45	0	1.03	1.05	0.43	—0.05	
3	F	—	U	—0.45	—0.35	—0.05	0.05	0	—0.20	—0.20	0.25	0	
4	F	—	B	—3.00	—2.50	—1.00	—0.85	0	—1.92	—1.40	1.08	—0.48	
5	F	—	B	—3.20	—3.05	—1.30	—1.10	0	—2.15	—2.20	1.05	—0.05	
6	F	—	U	—0.90	—0.75	—0.05	0.10	0	—0.40	—0.40	0.50	0	
7	F	—	B	—3.90	—3.60	—0.65	—0.20	0	—2.05	—2.20	1.85	—0.08	
8	B	0.7	U	—0.60	—0.40	0.25	0.40	0	—0.10	0.00	0.50	—0.20	Nearshore sediments, Visingsö island
10	B	2.7	U	0.50	0.90	1.80	1.90	<0.5	1.20	1.50	0.70	—0.43	
11	B	4.0	U	1.25	1.40	1.85	2.00	<0.5	1.63	1.65	0.38	—0.07	
31	B	12.0	U	1.25	1.40	2.10	2.25	2	1.75	1.75	0.50	0	
1960													
B 12	C	20	U	6.5	6.8	9.3	10.5	100	8.50	7.9	2.00	0.30	Glacifluvial sediment, Northern section
B 13	C	20	U	6.5	6.9	9.3	10.4	100	8.45	8.0	1.95	0.23	
B 16	C	20	?	7.9	8.7			100					
B 24	C	20	U	1.6	1.8	2.3	2.5	1	2.05	2.1	0.45	—0.11	
C 3	B	90	U	5.3	5.9	10.7		100		8.0			Deep trough sediment, Northern section
1961													
1	B	4.1	U	0.60	0.75	1.20	1.40	1	1.00	0.95	0.40	0.13	Nearshore sediments, Jönköping bay
2	B	2.3	U	1.25	1.45	2.20	2.40	3	1.83	1.80	0.58	0.05	
3	B	2.6	U	0.90	1.05	1.75	1.90	0.5	1.40	1.40	0.50	0	
4	B	3.0	U	0.75	0.90	1.55	1.70	1	1.23	1.25	0.50	—0.04	
5	B	2.5	U	1.00	1.10	1.70	1.80	<0.5	1.40	1.40	0.40	0	
6	B	2.9	U	1.00	1.15	1.70	1.80	1	1.40	1.40	0.40	0	
7	B	7.0	U	0.85	1.00	1.70	1.90	1.5	1.38	1.30	0.55	0.15	
8	B	1.7	U	1.50	1.65	2.30	2.50	4	2.00	1.95	0.50	0.10	
10	B	1.5	U	0.55	0.75	1.55	1.75	0.5	1.15	1.00	0.60	0.25	Nearshore sediments, Rosenlund cliffs
11	B	0.6	U	1.10	1.30	1.90	2.05	2	1.58	1.60	0.48	—0.04	
12	B	6.0	U	0.45	0.60	1.25	1.45	1	0.95	0.90	0.50	0.10	
13	B	1.3	U	0.30	0.45	1.25	1.50	1.5	0.90	0.80	0.60	0.17	
14	B	13	U	0.45	0.60	1.15	1.35	0.5	0.90	0.85	0.45	0.11	
15	B	1.7	U	0.75	0.95	1.85	2.10	3	1.43	1.35	0.68	0.12	
16	B	1.8	U	0.60	0.75	1.50	1.65	0.5	1.13	1.10	0.53	0.06	
17	B	2.1	U	0.55	0.70	1.20	1.35	1	0.95	0.95	0.40	0	
18	B	1.7	U	0.35	0.60	1.90	2.25	5	1.30	1.25	0.95	0.05	
20	B	1.1	U	1.65	1.85	2.70	2.90	13	2.28	2.25	0.63	0.05	
21	B	0.7	U	2.10	2.25	2.85	3.00	16	2.55	2.55	0.45	0	
23	B	6.4	U	0.35	0.75	1.85	2.15	4	1.25	1.30	0.90	—0.06	Nearshore sediment, Northern section
24	B	2.3	B	0.60	1.40	2.90	3.20	22	1.90	2.30	1.30	—0.31	Nearshore sediment, Rosenlund cliffs
26	B	11.8	U	1.30	1.60	2.35	2.60	8	1.95	2.00	0.65	—0.08	Nearshore sediments, Northern section
28	B	14.8	U	1.55	1.65	2.25	2.40	3	1.98	1.95	0.43	0.07	
32	B	0.6	B	0.00	0.65	1.70	1.90	2	0.95	1.35	0.95	—0.42	Nearshore sediments, Rosenlund cliffs
33	B	1.3	U	1.40	1.75	3.40	4.00	33	2.70	2.45	1.30	0.19	
34	B	2.3	U	0.90	1.15	2.50	2.90	14	1.90	1.80	1.00	0.10	
36	B	5.1	B	—3.70	—3.25	0.80	1.35	1.5	—1.18	—2.00	2.53	0.32	Nearshore sediment, Northern section
45	C	55	U	5.0	5.5	9.6	10.7	100	7.85	7.10	2.85	0.26	Deep trough sediment, Northern section
49	B	7.7	B	5.7	6.3	9.6	11.1	100	8.40	7.6	2.70	0.30	Nearshore sediments, Northern section
50	B	16.5	B	1.70	2.00	3.45	4.80	37	3.25	2.65	1.55	0.39	
51	B	6.4	U	4.80	5.10	6.90	7.80	100	6.30	5.85	1.50	0.30	
55	B	14.3	U	2.25	2.50	3.30	3.60	44	2.93	2.90	0.68	0.04	
59	B	19.0	B	2.90	3.35	5.25	6.05	83	4.48	4.20	1.58	0.18	Offshore sediment, Rosenlund cliffs

Continued

Nos.	Type	Depth	Mod.	Ø$_{16}$	Ø$_{25}$	Ø$_{75}$	Ø$_{84}$	%<3Ø	M$_ø$	Md$_ø$	σ$_ø$	α$_ø$	Notes
1961													
63	B	0.9	U	1.85	1.90	2.30	2.40	1	2.13	2.10	0.28	0.11	Nearshore sediment, Northern section
65	B	11.3	U	1.00	1.15	1.80	1.90	4	1.45	1.45	0.45	0	Offsh. sed. Rosenl. upper part
66	B	11.3	U	1.15	1.30	1.80	1.90	1.5	1.53	1.55	0.38	—0.05	lower part
68	B	0.7	U	1.85	2.40	6.80	8.30	66	5.08	4.10	3.23	0.30	Nearshore sediment, Rosenlund cliffs
69	B	15	U	1.55	1.70	2.80	3.20	20	2.38	2.15	0.83	0.28	Glacifluvial sediment, Northern section
71	F	—	?	10.2	10.6			100		(11.4)			Varved clay, Visingsö island
72	F	—	U	6.8	7.3	10.5	11.6	100	9.2	8.6	2.40	0.25	
73	F	—	?	10.0	10.4			100		(11.4)			
74	C	20	U	7.3	8.2			100		10.1			Glacifluvial sediment Northern section
75	C	20	U	8.6	9.6			100					
76	C	20	U	6.2	6.6	9.2	10.0	100	8.10	7.80	1.90	0.16	
77	C	20	U	6,5	7.1	10.4		100		8.5			
78	C	20	U	5.7	6.1	8.0	8.6	100	7.15	7.0	1.45	0.10	
79	C	44	U	8.7	9.8			100					Glacifluvial sediment, Central section
80	C	44	U	8.6	9.7			100		11.2			
81	C	44	B	6.6	8.1			100		10.5			
82	C	44	U	7.6	8.5			100		10.3			
83	C	44	U	7.1	7.9			100		9.7			
84	C	44	U	9.1	10.0			100					
85	C	44	U	8.2	9.0			100		10.6			
86	C	44	U	8.0	8.8			100		10.5			
87	C	20	U	6.4	7.1			100		8.9			Glacifluvial sediment, Northern section
88	C	20	U	7.9	8.9			100		10.5			
89	C	20	U	7.6	8.5			100					
90	C	20	U	6.3	6.8	10.3		100		8.1			
91	C	20	U	6.1	6.5	8.8	9.5	100	7.80	7.6	1.70	0.12	
92	C	20	U	6.8	7.5			100		9.3			
93	C	20	U	6.1	6.4	8.2	8.7	100	7.40	7.2	1.30	0.15	
94	C	20	U	5.5	5.8	7.5	7.9	100	6.70	6.6	1.20	0.08	
97	C	26	U	7.3	7.9			100		9.7			Offshore sediment, Northern section
98	C	26	U	6.7	7.4			100		9.3			
99	C	26	U	6.5	7.3			100		9.2			
100	C	26	U	6.3	7.0			100		8.9			
101	C	26	U	5.5	6.4	11.1		100		8.5			
102	C	26	U	5.7	6.4	10.9		100		8.3			
103	C	26	U	4.8	5.1	7.9	9.0	100	6.90	6.3	2.10	0.29	
104	C	26	B	4.0	4.2	6.8	9.1	100	6.55	4.8	2.55	0.69	
105	C	26	U	6.0	6.6	10.1	11.0	100	8.50	8.1	2.50	0.16	
106	C	26	U	6.7	7.6	10.9		100		9.5			
107	C	26	U	6.4	7.5			100		9.5			
108	B	2.3	U	5.20	6.25	10.4		96		8.35			Glacifluvial sediment, Rosenlund cliffs
109	C	26	B	3.7	3,9	7.3	9.3	?	6.50	4.4	2.80	0.75	Offshore sediment, Northern section
110	C	26	U	6.3	7.2	10.4	11.1	99	8.70	8.8	2.40	—0.04	
111	C	38	B	1.8	2.2	6.3	7.9	40	4.85	3.4	3.05	0.48	Glacifluvial sediment, Central section
112	C	38	U	7.7	8.6			100		10.3			
113	C	38	U	5.4	5.9	9.5	10.4	100	7.90	7.4	2.50	0.20	
114	C	38	U	5.0	5.5	8.8	9.5	100	7.25	7.1	2.25	0.07	
115	C	23	B	3.3	3.6	5.4	6.2	90	4.75	4.3	1.45	0.31	Offshore sediment, Jönköping bay
116	C	23	B	2.9	3.2	4.6	5.4	82	4.15	3.7	1.25	0.36	
117	C	23	B	3.0	3.4	5.3	6.2	84	4.60	4.3	1.60	0.19	
118	C	23	B	3.1	3.4	4.8	5.5	86	4.30	4.0	1.20	0.25	
119	C	60	U	5.1	5.9	10.2	11.1	100	8.10	8.0	3.00	0.03	Glacifluvial sediment, Central section
120	C	60	U	5.9	6.9	11.1		100		8.9			
121	C	60	U	4.8	5.0	7.0	8.2	100	6.50	5.6	1.70	0.53	
122	C	60	U	4.5	4.9	9.4	10.4	100	7.45	6.6	2.95	0.29	

Continued

Nos.	Type	Depth	Mod.	Ø_{16}	Ø_{25}	Ø_{75}	Ø_{84}	%<3Ø	$M_{\text{Ø}}$	$Md_{\text{Ø}}$	$\sigma_{\text{Ø}}$	$\alpha_{\text{Ø}}$	Notes
1961													
123	C	38	U	6.3	7.2	11.7		100		9.5			Glacifluvial sediment, Central section
124	C	38	U	6.4	7.2	11.5		100		9.2			
125	C	38	U	6.8	7.6	11.6		100		9.6			
126	C	38	U	7.2	8.2	12.3		100		10.3			
127	B	8.0	B	2.30	2.50	3.35	3.70	60	3.00	2.8	0.70	0.29	Nearshore sediment, Visingsö island
1962													
1	B	6.3	U	2.25	2.40	3.10	3.35	32	2.80	2.75	0.55	0.09	Nearshore sediments, Jönköping bay
2	B	8.5	U	2.60	2.95	4.40	4.75	74	3.68	3.75	1.08	—0.06	
3	B	9.4	U	2.75	3.00	3.70	3.85	75	3.30	3.40	0.55	—0.18	
4	B	10.3	B	0.60	2.15	3.55	3.90	44	2.25	2.85	1.65	—0.36	
5	B	15.5	U	3.05	3.20	4.35	4.85	86	3.95	3.65	0.90	0.33	
6	B	20.5	U	3.60	3.80	5.35	6.10	97	4.85	4.40	1.25	0.36	
8	B	15.0	U	3.70	3.90	5.20	5.75	99	4.73	4.40	1.03	0.32	
11	B	10.0	?	2.65	2.95	3.70	3.80	74	3.23	3.40	0.58	—0.29	
12	B	9.3	U	2.75	3.00	3.70	3.85	75	3.30	3.40	0.55	—0.18	
13	B	8.0	U	3.05	3.25	4.10	4.30	85	3.68	3.65	0.63	0.05	
14	B	4.2	U	2.30	2.45	2.85	3.00	16	2.65	2.60	0.35	0.14	
16	B	9.4	B	—3.60	—0.60	1.00	1.15	1	—1.23	0.60	2.38	—0.77	
17	B	10.5	B	0.80	1.25	3.45	3.65	49	2.23	2.25	1.43	—0.01	
18	B	17.7	U	3.10	3.40	4.70	5.20	87	4.15	4.05	1.05	0.10	
19	F	—	U	1.50	1.60	2.15	2.40	5	1.95	1.85	0.45	0.22	Glacifluvial sediments, Rosenlund cliffs
20	F	—	U	0.40	0.60	1.55	1.80	2	1.10	1.05	0.70	0.07	
21	F	—	U	2.50	2.80	4.25	4.60	69	3.55	3.55	1.05	0	
22	B	3.9	B	—2.25	—0.35	1.35	1.80	4	—0.23	0.60	2.03	—0.41	Nearshore sediments, Jönköping bay
23	B	6.1	B	2.85	3.30	3.75	4.10	82	3.48	3.50	0.63	—0.03	
24	B	4.9	B	2.35	2.45	3.10	3.40	30	2.88	2.65	0.53	0.43	
25	B	1.8	U	1.40	1.45	1.85	1.95	0.5	1.68	1.65	0.28	0.11	
26	B	6.5	B	2.30	2.75	3.70	4.00	73	3.15	3.30	0.85	—0.18	
27	B	12.5	U	3.00	3.10	4.25	5.00	84	4.00	3.45	1.00	0.55	
29	B	10.3	U	0.60	0.80	1.45	1.70	3	1.15	1.15	0.55	0	
30	B	13.6	U	3.95	4.25	5.90	6.55	98	5.25	4.90	1.30	0.27	
31	B	10.6	U	0.70	0.80	1.35	1.60	5	1.15	1.00	0.45	0.33	
32	B	11.3	B	0.80	1.00	4.75	5.25	39	3.03	1.45	2.23	0.71	
33	B	15.0	U	3.20	4.20	5.75	6.30	98	4.75	4.90	1.55	—0.10	
34	C	19.3	U	4.05	4.25	5.85	6.65	98	5.35	4.90	1.30	0.35	Offsh. sed., Jönk. bay upper part
35	C	19.3	U	3.90	4.15	5.55	6.25	98	5.08	4.70	1.18	0.32	lower part
36	B	2.5	U	1.80	2.00	2.65	2.80	9	2.30	2.30	0.50	0	Nearshore sediments, Jönköping bay
37	B	6.3	U	2.45	2.60	3.35	3.60	45	3.03	2.95	0.58	0.14	
38	B	9.0	U	2.80	2.95	3.70	3.90	72	3.35	3.30	0.55	0.09	
39	B	12.1	U	3.25	3.40	4.25	4.60	92	3.93	3.75	0.68	0.26	
40	B	19.6	U	3.90	4.15	5.60	6.25	99	5.08	4.70	1.18	0.32	Offshore sediment, Jönköping bay
41	B	11.7	U	3.30	3.50	4.50	4.90	92	4.10	4.00	0.80	0.13	Nearshore sediments, Jönköping bay
42	B	9.2	U	3.25	3.45	4.55	5.10	90	4.18	4.00	0.93	0.19	
43	B	3.6	U	1.35	1.55	2.50	2.75	11	2.05	2.00	0.70	0.07	
44	B	2.1	U	1.70	1.90	2.80	3.00	16	2.35	2.35	0.65	0	
45	B	2.0	U	1.75	1.95	2.80	3.05	17	2.40	2.35	0.65	0.08	
46	B	6.3	U	1.25	1.35	1.95	2.15	9	1.70	1.65	0.45	0.11	
48	B	4.7	U	0.65	0.80	1.35	1.50	3	1.08	1.05	0.43	0.07	
49	B	1.9	U	1.45	1.65	2.25	2.45	2	1.95	1.95	0.50	0	
51	B	5.3	B	4.05	4.65	7.20	7.80	90	5.93	5.95	1.88	—0.01	Glacifluvial sed., Rosenlund cliffs
53	B	13.3	U	0.45	0.50	1.30	1.55	1	1.00	0.55	0.55	0.82	Offshore sediment, Rosenlund cliffs
54	B	17.2	B	1.60	1.95	5.60	6.50	57	4.05	3.75	2.45	0.12	Offsh. sed., Rosenl. upper part
55	B	17.2	B	0.80	1.00	3.70	4.70	29	2.75	1.45	1.95	0.67	lower part
56	B	14.5	B	1.55	2.85	5.80	6.65	74	4.10	4.20	2.55	—0.04	Offsh. sed., Rosenl. upper part
57	B	14.5	U	0.30	0.50	3.10	4.50	26	2.40	1.30	2.10	0.52	lower part

Continued

Nos.	Type	Depth	Mod.	Ø$_{16}$	Ø$_{25}$	Ø$_{75}$	Ø$_{84}$	%<3Ø	M$_{Ø}$	Md$_{Ø}$	σ$_{Ø}$	α$_{Ø}$	Notes
1962													
58	B	4.5	U	1.50	1.70	2.50	2.70	7	2.10	2.10	0.60	0	Neareshore sediments, western shore, Southern section
59	B	6.0	U	0.20	0.40	1.45	1.90	7	1.05	0.80	0.85	0.29	
60	B	8.1	U	2.90	3.05	3.40	3.45	79	3.18	3.20	0.28	—0.07	
61	B	10.7	B	2.40	2.60	3.25	3.40	44	2.90	2.95	0.50	—0.10	
62	B	3.3	U	1.90	2.10	2.70	2.90	11	2.40	2.45	0.50	—0.10	
63	B	2.3	U	1.05	1.20	1.90	2.10	1	1.58	1.55	0.53	0.06	
64	B	2.1	U	1.15	1.25	1.70	1.85	0	1.50	1.50	0.35	0	
65	B	9.7	U	1.10	1.40	2.95	3.30	23	2.20	2.10	1.10	0.09	
66	B	12.2	U	1.05	1.50	3.50	3.80	39	2.43	2.60	1.38	—0.12	
67	B	1.2	U	1.65	1.80	2.40	2.55	4	2.10	2.10	0.45	0	
68	B	4.0	B	2.60	3.15	3.65	3.75	80	3.18	3.45	0.58	—0.47	
69	B	10.7	U	1.90	2.10	2.95	3.20	23	2.55	2.50	0.65	0.08	
78	B	21	U	2.00	2.15	2.95	3.35	24	2.68	2.45	0.68	0.34	Offshore sediments, easter shore, Central section
79	B	20	U	1.40	1.50	2.00	2.30	8	1.85	1.75	0.45	0.22	
80	B	20	U	1.25	1.40	2.00	2.20	5	1.73	1.75	0.48	—0.04	
86: 1	C	22	U	6.6	7.5	12.1		100		9.6			Glacifluvial sediment, Northern section
: 2	C	22	U	7.0	7.8	12.0		100		9.7			
: 3	C	22	B?	6.5	7.1	11.1		100		8.6			
: 4	C	22	B	6.5	7.5			100		10.5			
: 5	C	22	U	7.5	7.8	9.7	10.5	100	9.00	8.6	1.50	0.27	
: 6	C	22	U	7.2	7.6	10.3		100		8.7			
: 7	C	22	B	9.3	10.1			100		11.1			
: 8	C	22	U	7.5	8.1	11.0		100		9.4			
89: 1	C	22	U	5.9	7.0	11.1		100		9.3			Glacifluvial sediment, Northern section
: 3	C	22	U	6.0	7.0	10.8		100		9.0			
: 6	C	22	U	5.2	6.1	9.6	10.4	100	7.80	7.8	2.60	0	
: 7	C	22	U	5.7	6.3	9.8	10.9	100	8.30	7.6	2.60	0.27	
: 8	C	22	U	5.0	5.7	9.2	10.2	100	7.60	7.2	2.60	0.15	
: 9	C	22	U	6.5	7.4			100		11.4			
: 10	C	22	U	5.3	6.1	10.4	11.0	100	8.15	8.1	2.85	0	
: 11	C	22	U	6.8	7.4	10.6	11.3	100	9.05	8.9	2.25	0.07	
: 12	C	22	U	5.4	6.3	9.6	10.2	100	7.80	8.1	2.80	—0.11	
: 16	C	22	U	7.0	7.6	10.8	11.4	100	9.2	9.2	2.20	0	
: 20	C	22	U	5.7	6.3	9.0	9.7	100	7.70	7.7	2.00	0	
: 25	C	22	U	5.9	6.8	9.2	9.7	100	7.80	8.1	1.90	—0.16	

References

Special abbreviations

BEB	Beach Erosion Board, Corps of Engineers, Washington D.C.
BGIU	Bulletin of the Geological Institutions of the University of Uppsala, Uppsala.
GA	Geografiska annaler, Stockholm.
GFF	Geologiska Föreningens i Stockholm förhandlingar (Transactions of the Geological Society of Sweden), Stockholm.
GIU	Geografiska institutionen vid Uppsala universitet (The Institute of Geography at the University of Uppsala), Uppsala.
GJ	The Geographical Journal, London.
JG	The Journal of Geology, Chicago.
JSP	Journal of Sedimentary Petrology, Menasha.
LUGI	Geografiska institutionen vid Lunds universitet (The Institute of Geography at the University of Lund), Lund.
Medd.	Meddelande från (Publications of).
SMHA	Statens Meteorologisk-hydrografiska anstalt (The Meteorological and Hydrological Office of Sweden (before 1946)), Stockholm.
SMHI	Sveriges Meteorologiska och Hydrologiska Institut (The Meteorological and Hydrological Office of Sweden (from 1946)).
TAGU	Transactions American Geophysical Union, Washington D.C.

Current maps of the Vättern basin

Topographic maps on a scale of 1:100,000.
Generalstabens karta över Sverige, sheets: 35. Jönköping, 44. Hjo, 54. Karlsborg, 64. Askersund.
Economic maps on a scale of 1:10,000.
Ekonomiska kartan över Sverige (planimetric maps from aerial photographs and ground surveys).
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